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THE COURSE OF THE TENNESSEE RIVER AND THE PHYSIOGRAPHY OF THE SOUTHERN APPALACHIAN REGION

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ABSTRACT

The Tennessee River has been interpreted as a pirate stream. This idea has been rather generally accepted and used. There are opinions to the contrary, but they do not prevail since they have not been so widely published and are seldom taught. It is the purpose of the following article to present an explanation of the course of the river based on the geologic history of the region and obviating the element of stream capture.

The course of the Tennessee River is anomalous and has given rise to various conjectures concerning its possible previous courses and as to how its present position was established. The headwaters of the Tennessee are in the Appalachian Valley, and, since this region is well defined structurally and physiographically, it is natural to speculate as to why the river does not continue within it in a southwesterly direction to the Coastal Plains and the Gulf of Mexico. The name Appalachian River has been given to a stream which is supposed to have held such a position at the close of the Cretaceous cycle of erosion and in its lower portion followed approximately the present Coosa Valley. The existence of this hypothetical stream has been affirmed and denied.

Below Chattanooga the Tennessee flows westward through a gorge which has been described as having a youthful appearance.

Walden Ridge to the north of the gorge and Sand Mountain to the south of it are synclinal plateaus capped by Coal Measure rocks.

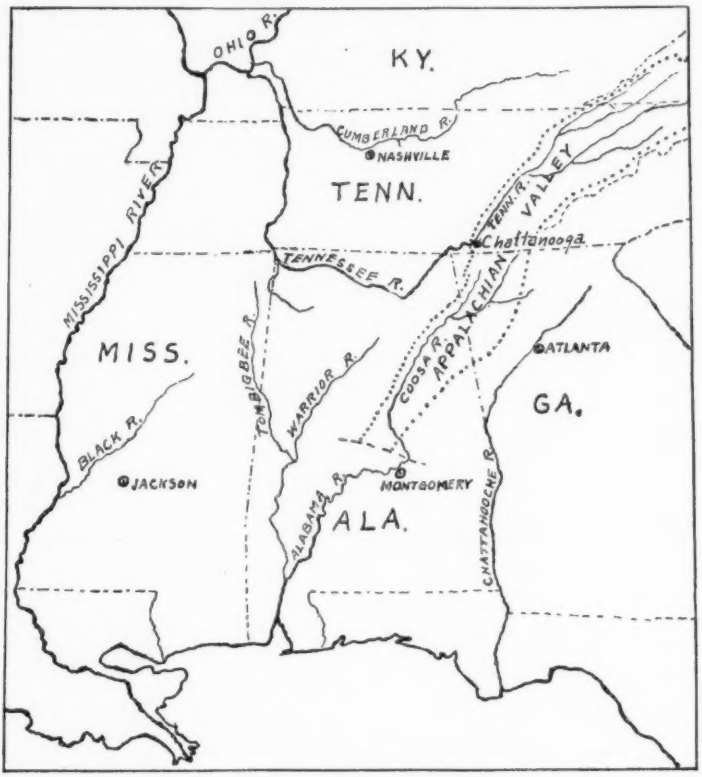

FIG. 1

It has been argued with the use of elaborate illustrations that the hypothetical Appalachian River[1] was captured at this place through headwater erosion by small tributaries of it and of a stream that

[1] C. W. Hayes, "Physiography of the Chattanooga District," *19th Ann. Rep.* (1899), *U.S. Geol. Survey*, Part II, pp. 9–58. See also, Hayes and Campbell, "Geomorphology of the Southern Appalachians," *Natl. Geog. Mag.*, Vol. VI (1894), pp. 63–126, for an interpretation which was later modified by Hayes.

flowed on the Sequatchie anticline and held a course in northern Alabama similar to that of the present Tennessee. It was reasoned, principally from physiographic evidence, that the diversion occurred in the Eocene, since which time the Coosa has been a decapitated stream and there has been a low divide in the Valley Region of northwestern Georgia on the northern portion of which northward-flowing tributaries of the present Tennessee have since been developed.[1]

The theory of an Appalachian River and the capture of its headwaters at the place of the Walden Gorge of the present Tennessee west of Chattanooga has been presented, with sketch maps to illustrate it, in one of our textbooks of geology,[2] and crustal warping or differential uplift at the beginning of the cycle of erosion which succeeded the Cretaceous, has been postulated as a contributing cause of the capture. The idea of river piracy is dramatic and its application to the Tennessee has gained wide credence, but it has been challenged, as will be explained.

In the textbook referred to, the stream, which is supposed to have captured the Appalachian River, is shown as flowing to the northwestern corner of Alabama along the course of the present Tennessee until it reaches the western state line, at which place it turns to the southwest and enters the Mississippi embayment of Eocene time in northeastern Mississippi. In the map illustrating a later condition, after the capture mentioned before, the stream in Mississippi is shown entering the Mississippi River by a course corresponding to the Black River, and the coast line is placed beyond the present border of the Gulf of Mexico. This is succeeded by a map in which an axis of uplift is indicated athwart the hypothetical stream in northeastern Mississippi. Deformation on this axis is supposed to have stimulated a tributary of the Ohio drainage to a capture and diversion which established the present course of the Tennessee River across western Tennessee and Kentucky to the Ohio River. It is a striking fact that the river flows northward from the northwestern corner of Alabama for a distance of about 150

[1] Hayes, *ibid.*

[2] Chamberlin and Salisbury, *Geology*, Vol. I, Advanced Course, pp. 164–68. (The maps used as text figures were based on articles previously cited and were drawn by E. S. Bastin.)

miles and approximately parallel with the Mississippi, but in a contrary direction, from a point where it is only about 100 miles distant from it. It would be very remarkable if this course was due to capture. The history of the Tennessee as thus interpreted is made doubly tragic.

The theory of the capture of an Appalachian river at the locality west of Chattanooga has been opposed by two writers. The first[1] reviews the evidence in favor of the theory and concludes that it is not sufficient. The second[2] goes into the subject at length and presents evidence to the contrary gained from an examination of the field.

It seems that the interpretation of the history of the Tennessee River has been based too largely on the study of the present drainage lines, and that inferences concerning their previous positions have been supported by questionable physiographic interpretations rather than from the geologic history of the region. It is rather useless to postulate courses of non-existent streams unless some evidence of their channels, or of deposits made by them, has actually been found. If the geologic history is correctly interpreted, the physiographic features will be in harmony with it. It is proposed to present here certain phases of the geology of the southern end of the Appalachians and the adjacent Coastal Plains which have a bearing on the course of the Tennessee and indicate their relations to the history of the Tennessee River.

The Paleozoic formations of the southern Appalachians suffered erosion during Triassic, Jurassic, and Lower Cretaceous times. The materials derived from them were transported so far that no evidence of stratified deposits of them is exposed in Alabama or encountered in the deep wells which have been drilled in the southern part of the state. This statement may probably be extended with perfect safety to include the states adjacent to Alabama. The area has generally been assumed to have been land in that time, and

[1] C. H. White, "The Appalachian River versus a Tertiary Trans-Appalachian River in Eastern Tennessee," *Jour. of Geol.*, Vol. XII (1904), pp. 34–39. This article was written in 1897 and refers only to the paper by Hayes and Campbell.

[2] D. W. Johnson, "Tertiary History of the Tennessee River," *Jour. of Geol.*, Vol. XIII (1905), pp. 194–231. This article refers to the later interpretation presented by Hayes.

it is natural to infer that the streams flowed away from the mountains, but no one can say positively just where the streams were,[1] and no one seems to have discussed any peneplains, or recognized any physiographic features which date from the Triassic or Jurassic.

The youngest formation which is found resting on the Paleozoic rocks in Alabama is the Tuscaloosa of Upper Cretaceous age.[2] It is a continental formation, at least for the most part and particularly where exposed. It has been reached by deep wells in the southern part of the state, and at some places where it has been passed through in drilling it has a thickness of 1,000 feet. It is a veneer of clays, sands, and gravels which were deposited on a plain that was somewhat diversified by escarpments, ridges, and anticlinal valleys. The character of the materials and the manner of their deposition give some idea of the direction from which they came and the sources from which they were derived.

The surface of the Tuscaloosa formation was an aggraded plain which was probably continuous with the valley floors of the streams which flowed from the Appalachian Mountains and crossed the areas of the Coal Measure rocks. The inland margin of the deposit was presumably irregular and well beyond its present limits. A small remnant of it has been found at a place halfway between the border of its main outcrop and Chattanooga,[3] and at an elevation which implies that it may have once extended farther in an irregular manner.

The character of the Tuscaloosa in Alabama changes along its outcrop. At the eastern margin of the state it is somewhat arkosic and contains materials derived from crystalline rocks. In the central parts quartz pebbles are conspicuous in its gravel beds. In the northwestern part of the state it consists in places largely of chert

[1] A paleogeographic map of Upper Triassic time accompanying an article by E. B. Branson on the "Triassic-Jurassic 'Red Beds' of the Rocky Mountain Region" (*Jour. of Geol.*, Vol. XXXV [1927], pp. 607–30) shows the Appalachian Mountains extending westward into Arkansas and Oklahoma, and a fictitious stream is indicated which has a part of its course corresponding to the position of the Tennessee.

[2] See the new geologic map of Alabama accompanying *Special Report No. 14, Geol. Survey of Alabama* (1926).

[3] George I. Adams, "The Occurrence and Age of Certain Brown Iron Ores of Alabama and Adjacent States," *Econ. Geol.*, Vol. XXIII (1928), pp. 85–92.

pebbles with a few quartz pebbles. This variation in the character of the materials implies that the drainages which brought them came from different areas of crystallines and Paleozoics. The quartz gravels in central Alabama presumably were brought from the Appalachian Mountains, while the chert gravels were derived from the Paleozoic limestones lying to the northwest of the Appalachian axis. It appears probable that the ancestral Tennessee River was the major stream of the drainage system of Tuscaloosa time that flowed southwestward into central Alabama. Its zone of deposition presumably shifted laterally from time to time and gradually retreated toward the northeastern part of the state as the invasion of the sea, which is recorded by the marine sediments which overlie the Tuscaloosa, advanced on the subsiding land.

The marine Upper Cretaceous has a thickness of over 1,000 feet in central Alabama and contains the Selma chalk, which is a clear sea deposit. It is reasonable to conclude that it once extended much farther inland than its present outcrop, and that at the time of the greatest transgression of the Cretaceous sea the ancestral Tennessee River which deposited Tuscaloosa sediments in central Alabama was shortened until its mouth or distributaries were well toward the northeastern corner of the state. If this is true, and inasmuch as the Tuscaloosa is a basal continental phase of the Cretaceous, I see no reason why the meanders which characterize the present Tennessee west of Chattanooga may not be inherited from an ancestral Tennessee river that had its distributaries a relatively short distance to the southwest of the present Walden Gorge. This interpretation obviates the necessity of discussing the question of an Appalachian river that was supposedly diverted across Walden Ridge to a stream on the Sequatchie anticline.

No attempt is made here to reconstruct the drainage over the pre-Cretaceous land surface. A plausible arrangement of the streams has already been shown by a sketch map illustrating the conditions at the middle stage of the deposition of the Tuscaloosa.[1] In it a stream is shown in the present position of the Tennessee River as

[1] E. W. Berry, "Upper Cretaceous Floras of the Eastern Gulf Region in Tennessee, Mississippi, Alabama, and Georgia," *U.S. Geol. Survey Prof. Paper 112* (1919), p. 27, fig. 7.

far as the Sequatchie anticline, and beyond that in the position of the headwaters of the Warrior River. This former position of the lower part of the Tennessee was postulated to agree with the interpretation of the Tuscaloosa as a deltaic deposit. It seems fully as reasonable to consider the formation as deposited on a sloping plain, in view of the fact that it has such a wide distribution and has been encountered in the deep wells in southern Alabama. According to this idea the stream may have shifted its channel very widely, and have once extended much farther than is indicated on the sketch map and well beyond the southern border of the present outcrop of the Tuscaloosa.

At the close of the Cretaceous deposition the southern Appalachian region and the adjacent Coastal Plains were elevated and probably warped. Following this uplift, the Tennessee River apparently took the position which it holds at present in its lower portion, by extending its course over the emerging land. It made a sharp bend near Guntersville and incised its channel in the unconsolidated Cretaceous sediments, and developed its valley in adjustment to the underlying Paleozoic rocks as they were re-exposed. In Alabama it parallels the present escarpment made by the Hartselle sandstone and follows the strike of the Paleozoics. There is some indication in the relatively wide portion of the Tennessee Valley of Alabama that it shifted its channel with the dip of the rocks. It seems wholly unnecessary to assume that it ever flowed across the state of Mississippi. No stream crosses the Cretaceous formations in Mississippi, or Tennessee, and it is reasonable to suppose that if one so important as the Tennessee had ever established such a course, it would have maintained it, since the Cretaceous rocks are easily eroded. The course of the Tennessee, where it flows northward, is with the strike of the Cretaceous formations, just as the course of the Tombigbee in general is with the strike of these same rocks in eastern Mississippi and western Alabama. The Tennessee River at the northwestern corner of Alabama, where it forms the state line, has a valley which is relatively narrow and like a gorge. It appears to have held this position since it established its course over the newly emerged Cretaceous deposits during the interval that preceded Tertiary deposition.

The foregoing explanation of the development of the Tennessee River is consistent with the commonly accepted interpretation of

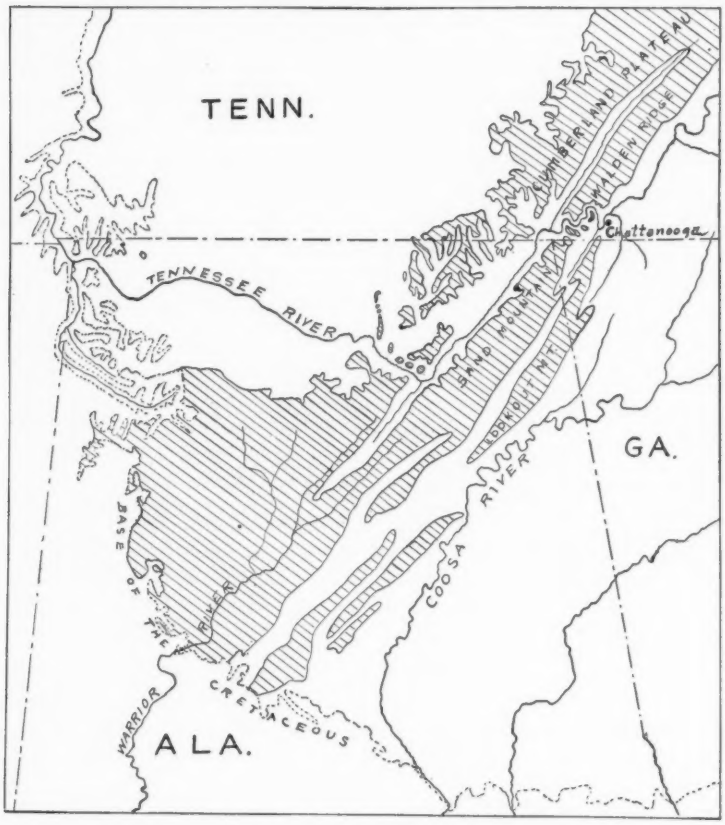

FIG. 2.—Section running northeast and southwest through Walden Ridge and Sand Mountain at the time of the greatest Cretaceous transgression.

the physiographic features of the southern Appalachians, based particularly on the study of the Chattanooga district. In what follows some suggestions are made in regard to the positions of the peneplains and base levels beyond the district and particularly in northwestern Alabama.

It is usually stated that the general surface of the Cumberland Plateau represents the Cretaceous peneplain. Walden Ridge is a part of this plateau, and the surface of the northern part of Sand Mountain, which is to the south of Walden Gorge, has been considered as representing the same peneplain. This mountain is a part of the area of the Coal Measures of Alabama which are shown by shading on the accompanying map, Figure 2. The Coal Measure areas are synclinal and are separated by anticlinal valleys. Toward the southwest they are diversified by minor erosional features and descend to lower elevations. It is a question as to just how far toward the southwest the Cretaceous peneplain may be recognized and as to where its limits may be placed. The accompanying sketch section, Figure 3, will assist in discussing this point.

FIG. 3

In this section five lines are shown. Three plains are represented which converge toward the Cumberland Plateau. The lower line represents the pre-Tuscaloosa land surface; the middle one, the upper surface of the Tuscaloosa formation; the third, the upper surface of the marine Cretaceous. The fourth line is the level of the Cretaceous sea, and the fifth line, the surface of the Cumberland Plateau.

The pre-Tuscaloosa land surface was the result of erosion during Triassic, Jurassic, and Lower Cretaceous times. South of the present belt of outcrop of the Tuscaloosa formation it is deeply buried. It is suggested in the section, in accordance with the interpretation of the Tuscaloosa formation offered before, that the Cumberland Plateau stood above the level of the plain on which the Tuscaloosa was deposited, and that its surface then corresponded approximately with the summits of what are now the residual areas[1] on the plateau. There were, presumably, anticlinal valleys and moderate escarpments on the pre-Tuscaloosa land surface, but they were much less

[1] Hayes, *ibid.*, Plate II.

pronounced than those now found in northeastern Alabama. The section shown in Figure 2 extending in a northeast-southwest direction lies in a shallow syncline and does not feature the anticlinal valleys which were filled by Tuscaloosa deposits.

If the Tuscaloosa is progressively younger toward the northeast, then the upper surface of the Tuscaloosa and the upper surface of the marine Cretaceous coincided at the level of the Cretaceous sea, and the surface of the marine deposits blended with the Cretaceous peneplain on the Cumberland Plateau. The Cretaceous peneplain is generally recognizable to the southwest of the Cumberland Plateau on the surfaces of the outliers of the Coal Measures to the west of the Tennessee River as far as the place where the river makes the sharp change in direction toward the northwest. It is also represented on Lookout Mountain, which is the easternmost area of the Coal Measures, and extends from Alabama into Georgia and terminates at Chattanooga. The general surface of Blount Mountain, which is a spur or lobe of Sand Mountain, is at such elevations as to correspond to it, but its high area rises above the peneplain. The surface of Sand Mountain, however, descends somewhat more rapidly toward the southwest, and its southern part, where it blends with the expanded area of the Coal Measures called the Warrior Basin, is so low that it probably was covered by the Tuscaloosa and possibly by the marine Upper Cretaceous deposits.

It is manifestly impossible to draw a line limiting the extent of the Cretaceous peneplain on Sand Mountain, but it may be inferred from a study of the present elevation and surface features of Sand Mountain that much of its southern part was reduced by erosion in Tuscaloosa time to lower levels than the Cumberland outliers and Blount Mountain.

The Intermediate Ridges of the Appalachian Valley in Tennessee and corresponding features in the Valley Region of northeastern Georgia and in the anticlinal valleys of northeastern Alabama have been interpreted as representing the Eocene base level, which has been named the Highland Rim peneplain. The Highland Rim in eastern Tennessee has elevations of 1,000 feet, and rises to the northeast. It is presumably at lower elevations in northern Alabama. Near Chattanooga the Intermediate Ridges are at elevations of about

1,000 feet. The 700-foot contour is at the border of the Tennessee, and the Eocene peneplain is about 250 feet above the flood plain of the river. In the Sequatchie Valley near the Tennessee River

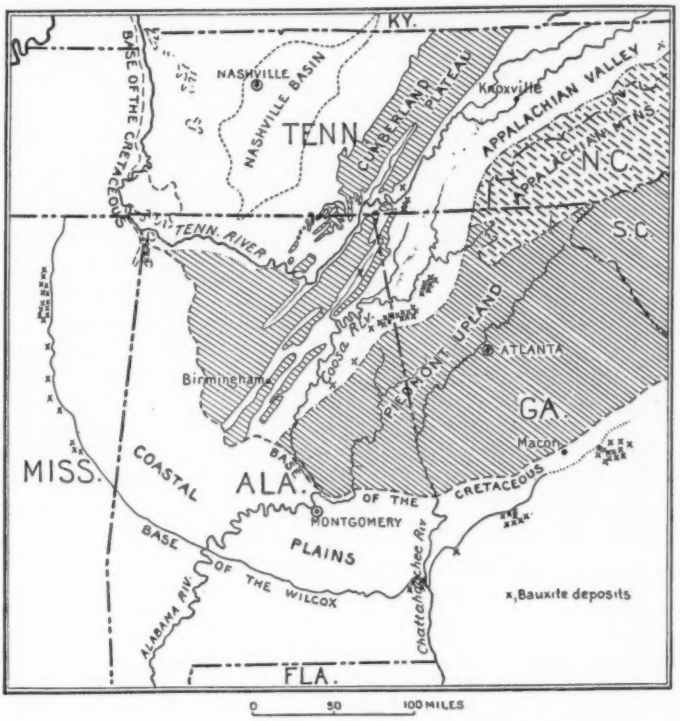

FIG. 4.—Bauxite deposits of the Southern Appalachians and adjacent Coastal Plains. By George I. Adams.

Tertiary gravels occur 230 feet above the flood plain.[1] The peneplain descends southward in the anticlinal valleys, and presumably also along the valley of the Tennessee.

There is reason for believing that the Eocene base level is recognizable in the northwestern corner of Alabama along Bear Creek, which is a tributary of the Tennessee River from the south just to

[1] Johnson, *ibid.*, p. 227.

the east of the state line. Well above the flood plain of this creek, which has a rather wide valley, there are bauxite deposits which occur in pockets having a considerable vertical range in the Paleozoics. They occur below the Tuscaloosa formation which is on the uplands. If, as has been suggested,[1] these bauxite deposits which are at elevations of between 500 and 600 feet are of the same age as those found on the Intermediate Ridges in the Appalachian Valley and on the anticlinal valleys of northeastern Alabama, they afford a means of identifying the Eocene base level in northwestern Alabama. The bauxite deposits on Missionary Ridge in the eastern part of Chattanooga are at an elevation of about 900 feet and are about 250 feet above the flood plain of the Tennessee. Those on Bear Creek are about 150 feet above the flood plain of the Tennessee River, which is at 400 feet at the mouth of Bear Creek. This difference in elevation of the bauxite at the two localities is consistent with the change in the grade of the stream in the intervening distance.

This interpretation of the Eocene base level is decidedly at variance with the suggestion that it is represented at an elevation of about 1,000 feet near Waynesboro, Tennessee, a town which is situated about 35 miles north of the mouth of Bear Creek.[2] Waynesboro is on the divide east of the Tennessee River in what has been called the western Highland Rim region of Tennessee. The Eocene base level has been called the Highland Rim peneplain because it is thought to be represented by the surface of the eastern Highland Rim region of Tennessee, but it does not necessarily follow that the western Highland Rim region of the state represents the same physiographic stage.

The 1,000-foot level near Waynesboro represents the re-exposed and modified pre-Tuscaloosa land surface, since remnants of the Tuscaloosa formation are found at intervals on the divide east of the Tennessee River from the Alabama line northward into Kentucky. The contact between the Tertiary and Cretaceous in western Ten-

[1] George I. Adams, "Bauxite Deposits of the Southern States," *Econ. Geol.*, Vol. XXII (1927), pp. 615–20.

[2] E. W. Shaw, "The Pliocene History of Northern and Central Mississippi," *U.S. Geol. Survey Prof. Paper 108* (1917), p. 52.

nessee is west of the Tennessee River, and the Cretaceous peneplain was once represented above the divide on which Waynesboro is situated. Presumably it extended eastward to the Cumberland Plateau, but it is extremely doubtful whether or not the Cretaceous formations were deposited throughout this whole distance, although they once extended well beyond their present area. A peneplain is not necessarily devoid of topographic features, and the western escarpment of the Cumberland Plateau may have been conspicuous in pre-Tuscaloosa time, and the eastern Highland Rim, and particularly its northern part, may have been a plain of degradation during the deposition of the marine Cretaceous.

The Coosa peneplain which has been assigned a Neocene age and lies below the Eocene base level has been recognized along the Tennessee River in the Chattanooga district, but is rather indefinite toward the western border of Alabama and has no topographic expression in the narrow river valley at the state line.

In conclusion: The suggestion is offered that the history of the Tennessee River, as outlined before, is relatively simple and has the merit of being in accord with the generally recognized geological history and structure of the regions through which it flows. The physiographic features of the Chattanooga district are readily interpreted without postulating any modification of the general course of the stream within that area. The presence of bauxite deposits in the valley of Bear Creek in northwestern Alabama is evidence of the position of the Eocene base level there with respect to the river, and is contributary proof that the course of the Tennessee where it flows in a gorge-like valley and forms the Alabama state line has been maintained since the close of Cretaceous time.

GEOLOGY OF BRITISH HONDURAS

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ABSTRACT

British Honduras consists of a central peneplain of folded Upper Carboniferous marine beds, with granitic intrusions, surrounded by unfolded limestone of about Oligocene age. Deposits between these two periods are unknown. The main movements are associated with Permo-Carboniferous and Miocene diastrophism. The trends of the Paleozoic rocks conform to those of Honduras and Guatemala. Youthful NE.–SW. features parallel to western Cuban trends are common. Large faults descending into 2,000-fathom water outline the east coast.

British Honduras (area, 8,210 square miles) is the only British possession on the mainland of Central America, and until recently was perhaps even less known than the adjoining republics. The geological survey was commenced in 1921, but as minerals in workable quantities were not located, it was abandoned at the end of 1926. The only previous geological investigations of the colony were G. H. Wilson's[1] geological traverses of rivers in 1886, and Dr. Carl Sapper's[2] traverses in the dry seasons of 1891, 1894, and 1896.

For the present purpose the Caribbean region may be defined as the area bounded on the northwest by the Isthmus of Tehuantepec, on the northeast by the Bahama Banks, and on the south by the Orinoco River.

The European idea of the Antillian Continent is that advanced by Suess.[3] The clearest statement of the case, however, is that by Dr. J. D. Falconer[4] in the *Scottish Geographical Magazine* (1902). Briefly, Suess's idea was that the Andes, on reaching Columbia, turn eastward along the north coast of South America through

[1] G. H. Wilson, "Notes on River Surveys, 1886," *Local Publication, British Honduras*.

[2] C. Sapper, "Über Gebirgsbau und Boden des nördlichen Mittelamerika," *Petermanns Mitt.*, Ergänz. 127 (1899); "Südlichen Mittelamerika," *ibid.*, 32 (1905); "Geology of Chiapas, Tabasco and Yucatan," *Jour. of Geol.*, Vol. IV (1896), p. 938.

[3] E. Suess, *The Face of the Earth*, English translation by H. B. C. Sollas, 1909.

[4] G. D. Falconer, "Evolution of the Antilles," *Scottish Geog. Mag.*, Vol. XVIII (1902) p. 37.

Trinidad, then turn sharply northward through some of the Leeward Islands, then sharply again westward through Jamaica and Honduras, and thence northward up the Pacific Coast. Suess originally divided the Antilles into an Inner zone (most of the Lesser Antilles), an Intermediate Zone (Caribbean Andes and Greater Antilles), and an Outer Zone (Florida and the Banks). He afterward considerably modified this idea. Exponents of this Antillean Continent theory omit to state clearly when it was formed and when it commenced to break up.

The Americans put forward quite different ideas. Bailey Willis,[1] in his paleographic maps of North America assumes that there have been Gulf of Mexico and Caribbean "deeps" ever since early Paleozoic times. Schuchert[2] suggests the existence, during Permian to Jurassic times, of the land mass of "Columbia," embracing the Greater Antilles, Guatemala, and Honduras, with an extension into the Pacific. T. W. Vaughan[3] has divided the region between the two Americas into twelve zones, but does not suggest that they all formed a continental mass. The author's own ideas of the problem favor the American views.

In the Caribbean region rocks definitely older than Carboniferous are unknown. Sapper's[4] maps show "Archean schists"; Sidney Powers[5] and myself agree that the granitic intrusions and the great earth movements at the end of the Carboniferous period could account for the metamorphism. The Lower Mesozoic is represented only by patches of limestone in western Cuba and fresh-water deposits in Honduras and Mexico. The Cretaceous is very widely distributed and exists throughout almost the whole of the Carribbean region, though often masked.

[1] B. Willis, "Index to Stratigraphy of North America," *U.S. Geol. Survey Prof. Paper 71* (1912).

[2] C. Schuchert, "North American Geosynclines" (presidential address), *Bull. Geol. Soc. Amer.*, Vol. 24 (1923), p. 158.

[3] T. W. Vaughan, "Geologic History of Central America and West Indies," *Bull. Geol. Soc. America*, Vol. XXIX (1918), p. 615.

[4] C. Sapper, "Über Gebirgsbau und Boden des nördlichen Mittelamerika," *Petermanns Mitt.*, Ergänz. 127 (1899); "Südlichen Mittelamerika," *ibid.*, 32 (1905); "Geology of Chiapas, Tabasco and Yucatan," *Jour. of Geol.*, Vol. IV (1896), p. 938.

[5] S. Powers, "Notes on the Geology of Eastern Guatemala," *Jour. of Geol.*, Vol. XXVI (1918), p. 507.

The diastrophism at the close of the Cretaceous appears to have been pronounced, except in Central America. Vaughan claims extensive submergence of Central America and West Indies in later Eocene and Oligocene times, with maximum submergence in mid-Oligocene. The Miocene was a period of great earth movements, and these are apparently still active. British Honduras is now outside the severe earthquake zone. The mountainous portion of British Honduras appears to have risen as a block and shows no folding since the intrusion of the granites. It has not been involved in the building of the present Central American mountain system, being situated on the lower side of its main arc. Very briefly, British Honduras may be described as a central peneplain of Upper Carboniferous rocks surrounded by Tertiary limestones.

TOPOGRAPHY

The portion of the Colony north of Lat. 17° 10′ is a limestone region of low relief. It contains the main watercourses, but at present is of little geological interest. The Belize and Sibun rivers collect the drainage from the Maya Mountains. The New River and Rio Hondo are base-level streams rising in the limestone lagoons.

The mountain area, known as the Maya Mountains, is a well-vegetated peneplain of about 3,000 feet altitude. The divide between the eastward and westward drainage is about 25 miles from the coast and about 16 miles beyond the western edge of the coastal plain.

The general east and west strike of the Paleozoic rocks has determined the easterly and westerly flowing streams. Although those trending eastward are little better than mountain torrents, they have not captured any important westerly drainage. All the colony's drainage enters the Caribbean except perhaps some small streams in the southwest corner of the Maya Mountains that disappear into limestones and perhaps form portions of the headwaters of the Rio Usumacinta.

The southern end of the colony contains badly fractured, thin-bedded mudstones and hills of massive limestone. The Sarstoon River forms the south boundary of the colony, and beyond it rise the limestone ranges of Santa Cruz.

The western boundary of British Honduras runs approximately north and south along Long. 89° 10′ W., but the southern or moun-

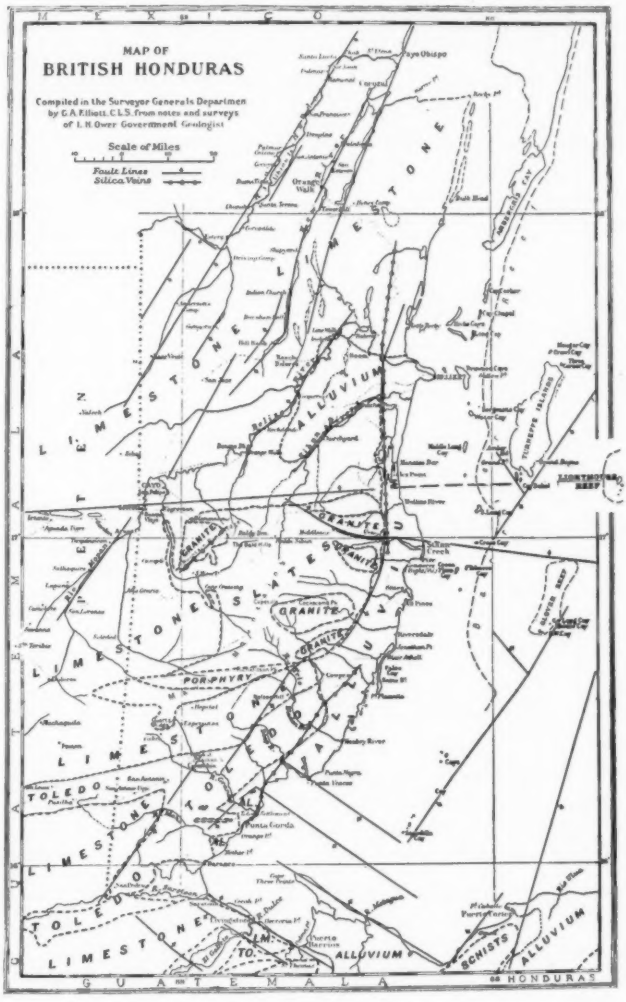

FIG. 1

tainous portion is still unopened. There is a coastal plain in few places exceeding 8 miles in width between the mountains and the sea. The levels of the Stann Creek railway indicate that the inland edge of the plain is 100 feet above sea-level. A trial railway survey run from Belize to Cayo gives 200 feet as the bank level at the junction of the Belize River and Rio Mopan. The highest point in the colony is the quartzite Cockscomb Peak, 3,800 feet, standing isolated a little to the seaward of the main divide. A number of ridges of the Maya Mountains exceed 3,000 feet in altitude.

The triangulation of the colony has just commenced, but when the geological survey began there was no map sufficiently accurate even for sketch geology. The local topographical draughtsman has compiled an entirely new map of British Honduras and has included the topography obtained during the geological survey. This has enabled contours to be sketched and a geological model to be constructed.

GEOLOGY

For the purpose of the present survey the geological formations of British Honduras have been classified as follows:

	Square Miles	
Coastal alluvium	1,100	Post Pliocene
Toledo beds	650	Miocene
Rio Dulce limestones and chalky marls	5,070	Oligocene / Miocene
Acid igneous intrusions	370	Permo-Carboniferous
Slate series	1,020	Upper Carboniferous
Total area of mainland	8,210*	

* Of this, 36 per cent is below the 500-ft. contour and 57 per cent above 1,000 ft. in altitude.

The survey being mainly for minerals, the Paleozoic rocks received most attention. Fossils are very scarce. The blue crinoidal limestone at Rio Trio and south of the Bald Hills contain Upper Carboniferous fossils. Foraminifera from the massive limestone of low altitudes were in 1896 considered Eocene (?). Vaughan now considers the fossils from the Rio Dulce gorge as Lower Miocene.[1] The Toledo beds and the New River chalky marls have in several localities yielded Miocene fossils.

[1] J. W. Vaughan, "Anguilla Formation," *Bull. Geol. Soc. America*, Vol. XXXV (1924), p. 733.

DEPOSITS STILL FORMING

The development of extensive coral reefs off British Honduras is due to the presence of platforms favorable for coral growth, facing currents bringing a good food supply. There is a barrier reef and three outlying banks. Fringing reefs are absent, and there are no traces of reefs inland. Vaughan estimates the growth of Atlantic coral to be from one foot in 14 years to one foot in 50 years. He states that he is unaware of Miocene or Pliocene coral reefs in the Antilles, and considers that coral reefs have played an insignificant part in building up the islands and banks of that region. Agassiz makes a somewhat similar statement and adds that the Campeche Bank coral is underlain by marine limestone, while along the Honduras-Mosquito coast and north of South America, corals appear upon extensive banks or shoals of proper depth. The opinions of these authorities coincide with the author's observations.

THE BARRIER REEF

The edge of this reef runs parallel with the shore line at a distance of from 15 to 22 miles. There are few channels through it that are navigable, the principal deep one being at English Cay, 15 miles southeast of Belize, where there are depths up to 150 feet. The sides are evidently limestone, as attempts to drive piles for lighting buoys had to be abandoned. Between the southern end of the barrier reef (Sapodilla Cays) and Guatemala there is a stretch of deep water giving access to Belize. Soft limestone and coral can be seen on the bottom in the water north of Belize, as the area between the reef and the shore is very shallow. On the seaward side there is a steep drop off the reef into deep water, 600 feet being recorded off English Cay and 1,200 feet of Sapodilla Cays. Numerous small cays and sand banks are exposed along the course of the reef. The water inside the barrier reef deepens gradually southward and there are many small cays, banks, and submerged pinnacles. Shells contribute largely to their formation, but on the outer cays coral predominates.

The rock 20 feet below Belize on which the pile foundations of the post-office, bridge, and customs wharf rest, may be a silt-covered coral reef.

DEEP SEA REEFS

The dimensions of the British Honduras deep-sea reefs are:

	Square Miles
Glovers Reef (22×6 miles)	130
Lighthouse Reef (30×6 miles)	180
Turneffe Reef (37×13 miles)	480

Off these reefs there is a sharp drop into over 1,000 feet of water. They lie with their long axis representing the continuation of the trend lines of a large portion of the barrier reef, which itself runs parallel with both the shore and the edge of the hills.

The reefs are mainly just below sea-level, but there are patches of dry sand formed by storm beaches. At Turneffe these sandbanks are numerous and are fringed with mangrove. In 1920 an oil seepage was reported from Long Cay on Lighthouse Reef, but it has not reappeared since.

ALLUVIAL DEPOSITS

These consist almost entirely of the deposits of the old coastal plain, ferruginous sandy clay predominating. Alluvial deposits are best developed between Deep River and Belize, but no idea can be formed as to their thickness or the nature of the underlying rocks. North of the Belize River the drainage is all from limestone, so little alluvium is deposited. In the hilly area the valleys are too steep and narrow to retain much detritus, but in a few localities when crossing granite they open out into fertile basins. From Punta Gorda to Swasey Stream there is a line of young siliceous veins and mounds having the usual 215° trend.

FLINTS

About nine miles inland from the mouth of the Sibun River flints appear on the alluvium and run northward. From Baker's Pine Ridge to Northern River, flints up to 2 feet in diameter are numerous. Beyond this, in the northeast corner of the colony, flints are frequent and quartzose veins traverse the limestone. The whole of the sand of the pine belts appear to have originated from siliceous springs, those that were submarine apparently also provided silica for the flints. Sapper reports that flints extend into Yucatan to the west of Rio Hindo. Round the township of Cayo flints are numer-

ous in a chalky marl which overlies the limestone at an altitude of about 600 feet. On the edge of the Little Mountain Pine Ridge, a few miles to the southeast, are young deposits of quartz grit. The siliceous springs that produced this grit may have also provided silica for these Cayo flints and sand for the Mountain Pine Ridge. In only two places were flints noticed in the solid limestone, and both are near quartz veins that may be contemporaneous with the limestone.

From Monkey River to Sarstoon River there are a number of warm springs, but only at the north end in Sennis Creek is siliceous material still being deposited.

CHALKY MARLS

These are confined to the north of Lat. 17°, and are best developed at Cayo, where NE.–SW. faulting has elevated them 600 feet above sea-level. Orange Walk and Corozal townships on the New River are also built on this marl, which has not yet been observed interbedded with the limestone.

THE TOLEDO SERIES (MIOCENE)

Between the Maya Mountains of British Honduras and the mountains of Guatemala are a series of thin-bedded strata to which the name Toledo beds has been applied, as in British Honduras they are confined to that district. Their weathering, with the heavy rainfall, produces the most fertile soil in the colony. Interbedded with these thin-bedded shales and mudstones are laminated limestones and blue calcareous Miocene sandstones. Diatomite occurs in Golden Stream and Temash River. Generally speaking, the farther from the sea, the fewer the calcareous bands, the less disturbed the strata, and the more unconsolidated the formation. The Toledo beds probably rest conformably on the Rio Dulce Limestone, but no section has yet been seen showing them actually in contact with undoubtedly older rocks. Near the coast the beds are badly crushed and the folding is very irregular.

Exposures of Toledo beds are numerous beneath the alluvium in both Golden Stream and Deep River, where they are either level-bedded or have a small southerly dip. In both streams they cease

suddenly against the massive limestone where warm water flows in from small tributaries. In Golden Stream the last exposure seen is level-bedded, but in Deep River the maximum dip, viz., 20°, with strike at 160°, is recorded close to the limestone. In the lower reaches of all the southerly streams no rock is exposed. On Monkey River no rock is seen below the junction of Bladen and Swasey branches, but Toledo beds appear at intervals up to Trio Junction. Up Rio Trio no rock is seen until close to the hills, when a small outcrop of Toledo occurs, followed by a thin strip of massive limestone lying against the slates and porphyries. A small outcrop some miles below Cowpen forms the northern limit of exposed Toledo. Alluvium extends from here to the slate at the first fall of Swasey Stream.

To the south the Toledo beds are downfaulted against the Lower Miocene limestone hills of Santa Cruz, Guatemala, and to the east they disappear beneath the sea. To the northwest the limestone rises beyond them, but the Toledo beds enter Guatemala and may even extend to the southwest corner of Peten.

At a couple of miles south of Pocton, where the pine belt ends, there are beds of sand amongst the limestone. At Santa Toriba, farther north along the main trail to Peten, mudstone and pink and green sandstone are interbedded with the limestone. The dips measured averaged 20°, with the 220° strike so frequent amongst the newer formations.

RIO DULCE LIMESTONE

There is no limestone definitely older than that of the Rio Dulce gorge at Livingstone from which this formation derives its name. Fossils collected by Sidney Powers in 1917 from the walls of this gorge were mainly corals and oysters which T. W. Vaughan refers to the Emperado limestone of the Panama Canal Zone. Spines and fragments of echinoids on weathered rock surfaces from Fuerto San Filipe (Lake Isobal) are classed as Lower Miocene.

With the exception of the Paleozoic rocks of British Honduras, the whole of the peninsula of Yucatan contains Tertiary limestone. Sapper states that the greater part of northern Chiapas is Cretaceous limestone and dolomite, but as he maps the formations of the Rio Dulce gorge as possibly of that age, much of his Cretaceous limestone may be younger, expecially as he admits that nearly all

his fossils were derived from the southern part of the Cretaceous belt. Sapper also states that in Chiapas and southern Tabasco limestone is an unimportant member. The eastern extension of these Chiapas Tertiaries is represented in British Honduras almost entirely by limestone. This British Honduras limestone is mainly foraminiferal and weathers easily. It is white to cream in color, frequently recrystallized, and dissolves almost entirely in dilute hydrochloric acid. In no place does it attain an altitude of over 2,500 feet.

The massive limestone in the mountains is in most places horizontal and rests directly on the granite and the upturned edges of the Upper Carboniferous. At Cane Crossing, where the Belize River commences to fork, the limestone is 1,000 feet thick. As it is absent to the north, thin to the south, and does not extend far to the east, it is evident that the floor on which it accumulated was uneven.

The relation of the Rio Dulce limestone to the Toledo beds is obscure, the best evidence being at the Rio Jolocte on the frontier, where the Toledo beds are seen dipping away from the uplifted limestone. At Cattle Landing, Punta Gorda, the Toledo shales overlie a massive limestone which is mapped Rio Dulce, like the prominent conical hills rising out of the adjacent coastal plain.

THE INTRUSIVE ROCKS

There are six main outcrops of acid igneous rocks:

		Square Miles
Granite (290 sq. mils)	Mountain Pine Ridge	80
	Middlesex	56
	Silk Grass Creek	27
	Waha Leaf Creek	27
	Cockscomb	100 (approx.)
Porphyry	Rio Bladen	80

The boundary of the Cockscomb granitic area was not located during the survey, and there is evidently some granite to the east of the Bald Hills, for the Sibun River gravels below these contain granitic material. All the igneous rocks appear to be from the same granodiorite magma. Some are interbedded, but the majority were intruded into the Upper Carboniferous sedimentary beds soon after their deposition.

As a small quantity of tin was located, the Middlesex area was mapped in some detail and petrological examinations were made by Dr. A. Brammall at the Royal School of Mines, London. This area being at the junction of two main faults, it is not surprising to find the igneous rock much crushed. Specimens from Macaroni Hill above the Stann Creek limestone quarry are described as remarkably crushed and mylonitized biotite granite. The highest degree of metamorphism noted is where Mullins River enters the coastal plain. At the western end of the intrusion and close to the main east-west fault are aplites, pegmatites, and altered hornblende gneisses.

Washings from the creeks draining this area were separated by bromoform and yielded anatase, andalusite, brookite, chiastolite, epidote, garnet, ilmenite, kyanite, magnetite, molybdenite, monazite, rutile, sphene, staurolite, tourmaline, and zircon.

From the western end of the Mountain Pine Ridge, aplites and altered igneous gneisses are mentioned. A specimen from Silk Grass Creek was defined as a biotite granite. No petrological examination has been made of granitic areas drained by Swasey Stream.

Immediately south of the Swasey the igneous rocks change entirely, the granite ends, and the quartz-feldspar porphyry area begins. The only porphyry dyke noticed to the north is at the extreme northwest corner of the Mountain Pine Ridge. The ridge between Swasey and Rio Trio has not been traversed, so it is uncertain what relation the porphyry of Rio Trio has to the Cockscomb granite. Petrological examination of a specimen from Rio Trio is described as quartz porphyry or altered rhyolite, while the rock of Swasey Stopper itself may be a crushed andesite or diorite. A large mass of quartz-feldspar porphyry is exposed on Rio Bladen and extends westward through some of the highest parts of the Maya Mountains. Sapper is probably correct in mapping it as extending almost to Dolores. Porphyry is also seen at the head of Austral Creek and in several gullies off the Quartz Ridge. This "porphyry" embraces intrusions, dikes, sills, flows, and probably ash beds. The country being rough and uninhabited and carriers difficult to obtain, no detailed mapping has been attempted.

UPPER CARBONIEROUS STRATIFIED ROCKS

These rocks are exposed over an area of about 1,020 square miles and consist mainly of schists and slates. Sandstones are few and there is a thin bed of blue limestone. The whole of the series is much crushed, but metamorphism is only severe near the contact with intruded rocks. The region of greatest metamorphism is near Mullin's River, where chiastolite slates and andalusite schists are recorded. As a contrast, the dark-blue slates near the sources of the Rios Chiqubal and Trio are within a few degrees of horizontal.

The usual slate dip is about 65°, the direction being determined by its position in relation to the intruded igneous rocks. The prevailing strike is 250°. These rocks are confined to the mountain area, and their succession has not been worked out, but the thickness must run into some thousands of feet. The eastern extension of the slates and porphyry of the Quartz Ridge is seen at Long. 88° 50′, and there is a small outcrop of slate on the edge of the coastal plain.

The blue crinoidal limestone has yielded the only fossil evidence, and forms a good horizon, but unfortunately has only been noticed at the Rio Trio and a point 4 miles south of the Bald Hills. Sapper, in his traverse from Cayo to the Cockscomb, mentions conglomerate and blue limestone only once each; but while descending to Deep River, conglomerate is mentioned several times. Where weathered it is difficult to distinguish from a shattered quartzite.

The Manatee-Cayo scarp marks the northern limit of exposed Paleozoic rocks in Central America.

MEXICO AND GUATEMALA

As the republics of Mexico and Guatemala adjoin British Honduras, brief mention of the formations across the borders is desirable. The limestone of northern British Honduras extends throughout Yucatan, but otherwise Mexico gives little assistance in elucidating the geology.

A geological map of Guatemala, including British Honduras, was published by Carl Sapper in 1899 and contains much useful information. His oldest rocks are the "Ancient Schists," but Sidney

Powers[1] and the author think that some represent only the Upper Carboniferous series metamorphosed by the granitic intrusions. The beds strike east and west, and associated with them, though marked younger by Sapper, are the great serpentine belts of the Caribbean streams. Serpentine was not seen in British Honduras.

The Bay Islands off the Honduras coast are mainly schist and represent the extension of the Sierra de Meredon. Utilla is the most westerly, and pre-Tertiary rocks are absent, the highest point,

FIG. 2.—Geological model of the Caribbean: (1) Paleozoic and intrusions; (2) Mesozoic (mainly Cretaceous); (3) Lower Tertiary; (4) Upper Tertiary; (5) Volcanic; (6) Submarine. Photo by Sweeting.

Pumpkin Hill (300 feet), being basaltic tuff containing fragments of coral reef. Nearby is Stuart Hill (170 feet), of olivine basalt capped by small coral reefs. Powers thinks that it was an active cone in quite recent times.

The Pacific slope of Central America is composed of the same volcanic ejecta that mask the high plateau, with volcanoes reaching up to 13,000 feet in altitude. Yucatan appears to have been free from vulcanism since Paleozoic times.

TECTONICS

British Honduras is the most northerly exposed portion of the Honduranian geosyncline. As the folds lie parallel to those of Gua-

[1] S. Power, "Notes on the Geology of Eastern Guatemala," *Jour. of Geol.*, Vol. XXVI (1918), p. 507.

temala and Honduras, they are considered to have been produced by the same series of movements at the close of the Paleozoic. Igneous intrusions follow the crests of these east-west folds and are more numerous on the faulted Caribbean side. The granite of the Mountain Pine Ridge has, in addition, NE.–SW. trends. About Lat. 16° 40′, Long. 88° 50′, the slates lie almost horizontal, so it is possible that Upper Paleozoic rocks underlying the limestone of Peten are little disturbed. It is suggested that the Maya Mountains formed a geanticline, the adjoining synclines to the north and south being now filled with limestone. This could explain the scarp along the edge of the slates at Lat. 17° 5′, but from features in Peten and soundings northeast of Glovers Reef, an east-west fault is assumed. There may have been a syncline between the Maya Moun-

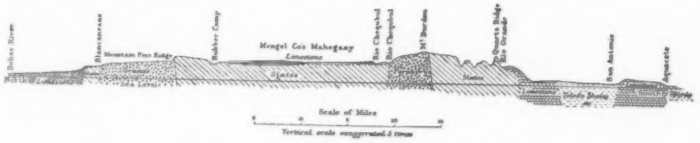

FIG. 3.—Section across British Honduras. Cayo–Punta Gorda

tains and those of Guatemala, but it is agreed that faults now mark the Miocene uplift to the south of the Rio Usumacinta.

Only Upper Carboniferous and Lower Miocene fossils have been identified from British Honduras; but as the latter are from low altitudes, the mountain limestone can be treated as Oligocene. Even if the sandstones interbedded with this limestone in southeast Peten are of hydrothermal origin, the associated mudstones indicate that the Maya Mountains were land surfaces at that period and that the drainage was westward. Serpentine grains in the Toledo series mean a Guatemala origin and a northeast drainage for that district. The denudation necessary to expose the granite now overlain by the Oligocene limestone could have taken place in Eocene times, but some Lower Mesozoic denudation is likely. The 3,000-feet Maya Mountain peneplain shows that the Paleozoic rocks were planed off to sea-level before the commencement of the Miocene uplift. On the southwest the slates of the Quartz Ridge rise rapidly 500 feet out of a 2,500-feet limestone plateau, but on the northeast metamorphic rocks rise out of dissected limestone of about 500 feet elevation.

On the frontier the limestone descends gradually in a northwest direction from the Maya Mountains to the source of the Rio Hondo, and is crossed by the prevailing NE.–SW. faults. In the vicinity of one of these faults, at the mouth of the Cayo gorge, is the only record of folded limestone. On the southeast corner of the Maya Mountains the limestone decreases in altitude until, at Swasey, it consists only of a few acres capping a granitic hill. The junction here follows the general strike of the slates. South of the Quartz Ridge there is a rapid fall, beyond which the limestone appears to merge into the thin-bedded Toledo series. The junction of the Toledo beds and the limestone to the south of San Antonio follows an east-west fault. Farther south the junction of the limestone with the coastal plain is a fault with the usual 210° trend. The limestone hills of the Toledo district coastal plain also trend in this direction. At the mouth of the Rio Sarstoon there is a small outcrop of Paleozoic(?) rock due to the uplift of the Livingstone limestone block which cut off El Golfeto from the sea. This Livingstone fault which is exposed at the mouth of the Rio Dulce gorge, trends 300°, or at right angles to the main fault system. The most important fault off the British Honduras coast lies to the east of Lighthouse Reef. Southward this can be traced to the Rio Ulua, and northward almost to the Yucatan Channel.

The Barrier Reef is diverted from the general northeasterly trend by right-angle bends toward the shore. Cross-faults are suggested; but owing to the lack of soundings, data are available only at the southwest corner of Turneffe. Soundings of 1,146 and 1,641 fathoms to the northeast of Glovers Reef, and 1,677 and 1,922 off the southeast corner of Lighthouse Reef, give a dislocation with a downthrow to the north. This may run westward into the Stann Creek Valley or die out to the south of Turneffe. Between Stann Creek and Belize the 210° trend lines become north-south features. With the exception of the Livingstone and the Rio Ulua faults, all features in Southeast Guatemala have the 250° Honduranian trend. As the coral does not rise above sea-level, the Barrier and outlying reefs belong to an area of subsidence. The coral probable rests on Tertiary limestone.

The long line of quartz and quartzite veins that traverse the allu-

vium, almost parallel with the coast, belong to the main fault system. Grits on the Mountain Pine Ridge that overlie the limestone at 1,400 feet elevation, and even cap the 3,000 feet Bald Hills are evidently derived from siliceous springs discharging on the high side of the faults. The thick quartz veins that form a prominent NE.–SW. feature in the Mountain Pine Ridge are of no great antiquity. Sand is contemporaneous with the limestone at the Pocton Pine Ridge, which is the southwest extension of this line, but it is equally probable that this silica may be connected with the east-west vein that forms the crest of the Quartz Ridge. Silicification even amongst the Paleozoic rocks appears to be of Tertiary age.

GEOLOGY OF DEEP SPRING VALLEY, CALIFORNIA

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ABSTRACT

Deep Spring Valley is one of the most remarkable desert basins in California. Several thousand feet of pre-Cambrian strata and 10,000 feet of Lower Cambrian strata are wonderfully exposed around portions of the valley. A granite-monzonite series, probably of late Jurassic age, intrudes the strata on a large scale.

Various conflicting ideas have been expressed in regard to the origin of the valleys of the Basin Range province. Deep Spring Valley very clearly owes its origin to faulting, the whole region being in fact a mosaic of normal fault blocks. Great fault facets are remarkably exhibited. The recency of the dislocation along the main fault is proved by many miles of sharply faulted alluvial cones; the occurrence of river beds on the top of a great fault block; and a field of very fresh lava which has been dislocated 1,500 feet, vertically.

In relatively late Quaternary time a deep fresh-water lake occupied the valley.

INTRODUCTION

The desert valleys of California abound in a great variety of significant geological features which are usually exhibited with wonderful clearness, though in many cases they are difficult of access. The writer recently spent several weeks in one of these fascinating valleys, called Deep Spring Valley, in and around which there is shown a combination of geological features extraordinary even among the remarkable desert valleys of California. The present paper embodies results of observations then made. It is a pleasure to acknowledge the many courtesies extended by Mr. O. B. Suhr and others connected with Deep Spring Ranch and School.

Deep Spring Valley is situated in the northern part of Inyo County, lying partly within the Bishop quadrangle, and partly within the Lida quadrangle. It is a northeasterly trending, oval-shaped basin with a floor 13 miles long and a maximum width of $4\frac{1}{2}$ miles. The altitude of the lowest part of this nearly flat-floored basin is almost 5,000 feet. The valley is completely surrounded by rugged mountains which rise 1,000–3,000 feet above its floor. There is no drainage outlet. The whole region considered in this paper covers about 135 square miles.

Knopf and Kirk,[1] in a reconnaissance report on the Inyo Range, include a map showing the distribution of the principal formations

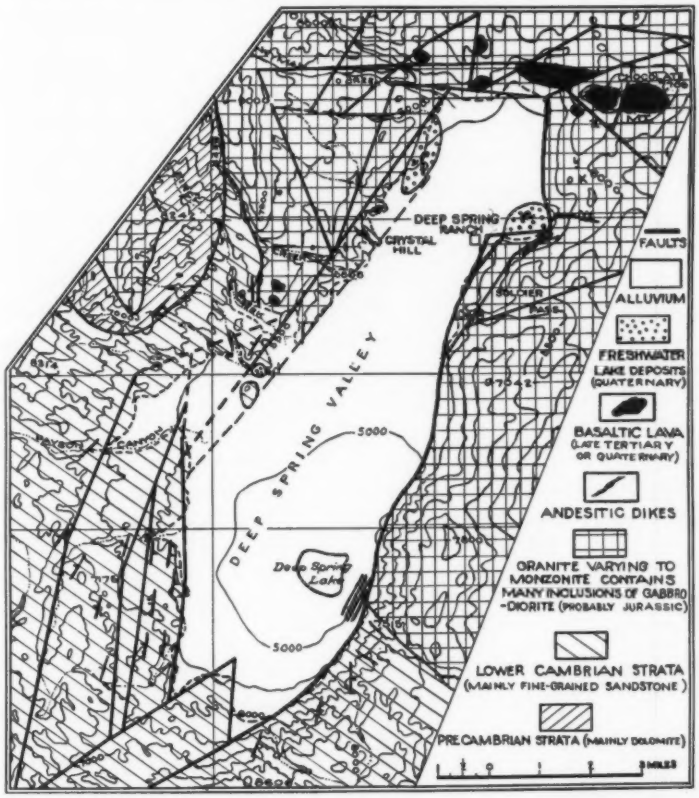

FIG. 1.—Geological map of the vicinity of Deep Spring Valley. Contours from United States Geological Survey maps.

in only the western portion of the Deep Spring Valley region, and they give considerable data on the stratigraphy of the Cambrian and pre-Cambrian. No fault lines are shown on their map, and little

[1] A. Knopf and E. Kirk, "A Geologic Reconnaissance of the Inyo Range and the Eastern Slope of the Southern Sierra Nevada, California," *U.S. Geol. Survey, Prof. Paper 110* (1918).

or no attention is given to various other important geologic and geomorphic features of the region.

THE ROCKS

Pre-Cambrian strata.—The oldest rocks of the region are believed to be pre-Cambrian. They are separated from undoubted Lower Cambrian strata by an unconformity. In the report already cited, Kirk says (p. 9) that these pre-Cambrian "rocks may be subdivided into three lithologic units—a series of sandstones and thin-bedded impure dolomites at the bottom, the Reed dolomite above these, and, locally, the Deep Spring formation at the top." The total thickness of these formations is about 3,600 feet. The southern end of a considerable area of these pre-Cambrian rocks lying northwest of Deep Spring Valley is represented on the accompanying map (Fig. 1).

Lower Cambrian strata.—Strata of this age are extensively developed to the west and south of Deep Spring Valley, as shown on the accompanying map. According to Kirk, the Lower Cambrian rocks are

> subdivided into the Campito sandstone which forms the base of the system, and the overlying Silver Peak group. The Campito sandstone, 3,200 feet thick, consists chiefly of dark, fine-grained, feldspathic quartzite with many argillitic and phyllitic partings. . . . . The Silver Peak group, 7,000 feet thick, consists largely of calcareous shales, sandstones, and limestone. Fossils occur in sediments of all three types.

Kirk made no attempt to separately delimit the Campito and Silver Peak on his map; nor has the writer on the map accompanying the present paper.

In addition to the extensive faulting of the Lower Cambrian strata, described below, these rocks are notably folded in a general north-northeasterly direction, as indicated on the accompanying map. Close to the contact with the monzonite on either side of the valley, the Cambrian strata show extra steep dips of 60°–80°, no doubt due to the shouldering action of the intruding magma.

Gabbro-diorite.—The oldest igneous rock of the region is a quartz diorite varying to gabbro-diorite. It occurs in many places, but nowhere in large bodies. Wherever found, it is in the form of very irregular inclusions ranging in size from less than an inch to half a

mile long. Many of these inclusions are sharply defined, while many others have been more or less intricately cut, injected, and digested by the granite. Excellent examples of the latter may be seen in Wyman Canyon. Myriads of sharply defined small inclusions occur among other places in the granite about half a mile up the canyon just southeast of Deep Spring Lake.

Diorite inclusions of all sizes are most common between Wyman Canyon and Birch Creek and from Soldier Pass southward for several miles. The largest observed inclusion is a body of very massive gabbro-diorite on the hill north of the mouth of Birch Creek. This rock contains, by volume, about 35 per cent each of microperthite and oligoclase; 20 per cent of augite; several per cent each of biotite and magnetite; and a little apatite and zircon. Most of the inclusions, however, are quartz diorite.

Granite.—The most common rock, by far, in the mountains which bound the valley on the northwest, north, and most of the east is a light gray, medium-to-fine grained granite varying toward granodiorite. Examination of a number of thin-sections shows this rock to contain the following minerals by volume percentages: microcline, 40–50; quartz, 30–37; oligoclase, 10–25; biotite or hornblende, 1–6; magnetite, 1 or less; titanite, 1 or less; and in some cases a little apatite or muscovite. This rock is generally non-foliated, and it exhibits little or no granulation.

The normal granite above described varies locally to hornblende or augite-rich syenite or quartz-monzonite facies. These more basic facies, in part at least, seem to have resulted from digestion or assimilation of the older gabbro-diorite by the granite magma. Thus, in Wyman Canyon Narrows and vicinity the older dark rock has been more or less intricately cut and digested by the granite. Many remaining inclusions of the dark rock there show all stages of absorption; and either a dark-gray biotite-rich, or pink hornblende-rich, rather massive quartz monzonite, carrying as much as 2 per cent of titanite, has resulted from the digestion.

One of the best places to observe a darker facies of the granite-monzonite series is on the great mountain face just east of Deep Spring Lake, where the rock is a gray, more or less massive, augite-rich syenite varying to monzonite with considerable amounts of

hornblende, magnetite, and biotite. This rock is believed to have resulted from assimilation of the old gabbro-diorite by the granite magma.

The granite in several places exhibits a remarkable large-scale type of jointing which may be called a platy structure. Thus the whole ridge hundreds of feet high, immediately facing the valley floor for more than a mile northwestward from the mouth of Birch Creek, consists of granite which, due to the excellence of this kind of jointing, may be mistaken readily for a great outcrop of uniformly

FIG. 2.—Steep-dipping joint plates in granite near the mouth of Birch Canyon

well-bedded, strongly tilted sandstone when viewed from a little distance (Fig. 2). There is also a remarkable exhibition of the platy structure northeast of Deep Spring Lake, where the granite in the precipitous mountainside 2,000 feet high, is split from top to bottom into a great series of rather evenly spaced, nearly vertical sheets extending into the mountain.

In regard to the age of the granite, all that can be said with certainty is that it is post-Cambrian and pre-Quaternary. Granite of presumably the same age cuts Triassic strata farther south in the Inyo Mountains, and this granite is probably of the same age as that of the Sierra Nevada Mountains. In the Deep Spring region dikes of this granite cut the Lower Cambrian strata.

Andesitic dikes.—Several of these dikes occur within the area, and they show considerable variation. All of them sharply cut the granite-monzonite series. Most interesting is the dike with a northwesterly trend sharply cutting the impure granite of the little ridge just northwest of Crystal Hill. Most of the dike material is a hard, ringing, light-gray, slightly porphyritic rock. Under the microscope it is seen to contain about 60 per cent of plagioclase (oligoclase to labradorite); 14 per cent of biotite; 25 per cent of very fine-grained, rather indeterminate groundmass; and a little pyrite, magnetite, and epidote. Southeast of Deep Spring Ranch, a more or less continuous, narrow, branching, dark-gray, andesitic dike several feet wide sharply cuts the granite along the general zone of faulting on the side of the mountain. A thin-section shows labradorite, biotite, hornblende, and secondary calcite in order of abundance. Due east of the ranch, a dike of similar material cuts across the mountain sharply in a nearly east-west direction.

Lava.—A number of fields of lava, shown on the accompanying map, occur around the northeastern end of the valley. It is more than likely that these were originally part of a single field which has been dislocated by faulting and partly removed by erosion. The main bulk of the rock is a dark-gray, fine-grained, olivine basalt containing the following minerals by percentage volume: andesine to labradorite, 45–60; augite, 20–40; olivine, 7–9; magnetite, 1–5; and in some sections a little indeterminate groundmass. Cellular, scoriaceous, and cindery facies are common on top of Chocolate Mountain. Two very small areas of similar lava occur on the hill just north of the mouth of Birch Creek.

There is no evidence that this lava was ever covered by later deposits of any kind. This fact, and the remarkable freshness of the rock, render it quite certain that the lava is not older than late Tertiary. More than likely it is early Quaternary.

GEOMORPHOLOGY

The system of faults.—Viewed broadly, Deep Spring Valley may be regarded as a trough-fault valley in a remarkable state of preservation. The valley is bounded by well-defined normal faults, and the mountains immediately surrounding the basin are broken into

a mosaic of small, normal fault blocks bounded by fractures running in various directions, as shown on the accompanying map. Various faults also quite certainly intersect the bed rock under the floor of the valley, but they are completely masked by the deep alluvial deposits.

The fault which bounds the valley along its whole eastern side may be called the "Deep Spring fault." It is marked throughout its

FIG. 3.—Great triangular fault facets northeast of Deep Spring Lake (R. H. Mansfield, photo).

length of at least 15 miles by a very bold scarp rising nearly everywhere from 1,000–3,000 feet above the valley floor. For a distance of 7 or 8 miles along its southern one-half, the fault scarp is the most wonderfully preserved large-scale phenomenon of its kind ever observed by the writer (Fig. 3). Great, triangular, precipitous facets, separated by short, very steep canyons, and rising hundreds of feet above the valley floor, in the vicinity of Deep Spring Lake, represent almost exactly the actual fault face very little affected by the weather. Above this great series of triangular facets, and rather sharply separated from them, the fault scarp has been more affected by weathering and erosion; and it is distinctly less steep. This indi-

cates that, after profound faulting had taken place, there was a distinct pause and erosion of the scarp, since which there has been a vigorous renewal of the process of dislocation extending to, or nearly to, the present time. The maximum displacement along the Deep Spring fault seems to be southeast of the lake, where it is at least 3,000 feet, the element of uncertainty being the depth of the alluvium in the valley at the base of the scarp.

A narrow, wedge-shaped, downthrown block of granite 2 miles long, with strong topographic expression, lies against the Deep Spring fault scarp west and south of Deep Spring Ranch. A similar, much smaller block lies against the larger block just back of the ranch.

Soldier Pass, which is the lowest place on the rim of the valley, has been eroded along a branch of the Deep Spring fault where it cross-cuts the mountain ridge on the east side of the valley.

The lava-capped mountain, locally known as Chocolate Mountain (altitude, 7,705 feet), which rises boldly above the northeastern end of the valley, forms part of a quadrilateral fault block, as shown on the map. The lava bed at the top of the mountain has been dislocated fully 200 feet by a small cross-cutting fault. The north face of Chocolate Mountain is a 35° slope corresponding closely with the exposed fault face at the base of the mountain. A feature of special interest is the relation of the fields of very fresh lava at the top and at the western base of the mountain. These lava beds were originally quite certainly part of a single field which has been dislocated by faulting to a maximum vertical extent of fully 1,500 feet (Fig. 4). The actual fault contact between the lower-level lava field and the granite at the base of Chocolate Mountain is very plainly visible, showing a dip of about 35°. Since the northwestern fault face of the mountain is almost as steep, it shows that little erosion has taken place since the great dislocation of the lava field. A very small area of similar lava, lying on the southwestern side of the Chocolate Mountain, has been down-faulted hundreds of feet from its former connection with the lava on top of the mountain. The fault features on and about Chocolate Mountain are remarkable exhibitions of such phenomena, being accentuated by the blackness of the lava and the barrenness of the mountain.

At the north end of the valley, near the mouth of Wyman Canyon, several other smaller patches of lava represent remnants of a once continuous sheet which has been dislocated and subdivided by small intersecting faults.

A series of four interesting, nearly parallel north-south normal faults, ranging in length between 5 and 7 miles, controls the major topographic features on the southwestern side of the valley (see

FIG. 4.—The slightly modified fault scarp 1,500 feet high forming the west face of Chocolate Mountain. The lava field on top of the mountain, the edge of which only is shown by the dark patch, was originally part of the lava field shown at the base of the scarp.

map). All the rocks in this area are stratified. The ridge, less than a mile wide and 5 miles long, which there forms the boundary of Deep Spring Valley, is a relatively uplifted fault block. There has been a displacement varying from a few hundred feet to more than 1,000 feet along the fault which separates this block from Deep Spring Valley. The southern half of this block has a very distinct anticlinal structure, and triangular fault facets are very well preserved along its eastern side. The northern portion of the block is irregularly much folded.

Separating the fault ridge just mentioned from another, larger,

more westerly fault block, there is a well-defined, narrow, fault-trough valley 5 miles long and several hundred feet deep. The western fault block just mentioned is a mile or more wide and at least 6 miles long. The northern half of this block has been downthrown fully 1,000 feet with reference to the mountain mass adjacent to it on the west. The uniform strike of the rather steep-dipping strata in the southern half of this block is parallel with the north-northeasterly trend of the block itself. From Payson Canyon south for 2 miles, the strata of this block are highly and irregularly folded, while north of the Canyon alluvial materials completely conceal the bed rock. All four of the faults just described end abruptly on the south against a northeasterly trending fault which has in part determined the position of a canyon.

A conspicuous, straight fault 9 miles long extends from the mouth of Payson Canyon to the mouth of Wyman Canyon along the northwestern side of Deep Spring Valley less than a mile from its margin (see map). The northern two-thirds of this fault extends through granite, and the rest of it through strata. Its maximum displacement is about 700 feet, with downthrow on the east. Topographic evidence points strongly to the existence of another long straight fault at the very margin of the valley and parallel with the fault last mentioned. Its eastern side is downthrown to a maximum of fully 700 feet. The total subsidence of Deep Spring Valley, with reference to the mountains on its northwestern side, is, therefore, at least 1,400 feet. Evidently, the important fault movements have taken place there much less recently than along the Deep Spring fault, not only because the fault scarps are not nearly so well preserved (no faceted spurs) but also because the alluvial cones and fans are there proportionately much better developed.

A relatively sunken fault block of granite, with the shape of a triangle, whose base is in Wyman Canyon and whose apex is near the mouth of Beer Creek, has moved down 1,500–2,000 feet relative to the adjacent areas of granite on the west and north. The topographic expression of this sunken block is very strong.

Recency of the faulting.—It is quite certain that Deep Spring Valley originated as a fault basin during Quaternary time. There is strong evidence that a considerable amount of the faulting must

have taken place during relatively late Quaternary time. There is, indeed, striking evidence of fault movements which have occurred within the last few centuries. Recency of the faulting is indicated by phenomena already mentioned, such as: great triangular fault facets; profound dislocation of a Quaternary lava sheet; fault scarps with topographic expression irrespective of kinds and structure of the rocks; and river beds on top of a high ridge (see below).

Another strong evidence that there has been much recent relative movement between the valley floor and the mountain ridge adjacent to it on the east lies in the fact that the alluvial cones at the base of the ridge are very small as compared to the amount of débris which has been carried out of the canyons. This is particularly true from Soldier Pass to south of the lake. Evidently the floor of the valley adjacent to the ridge has sunk so fast and so recently that the coarse alluvial material, even at the mouths of such steep desert canyons, has been carried down on the subsiding valley floor almost as rapidly as it has been washed out of the canyons. In connection with this phenomenon, it is significant that the lowest part of the valley (at the lake) lies adjacent to the highest, very steep fault scarp in the whole area.

Perhaps the most interesting phenomenon indicating great recency of considerable fault movements is the dislocation of the alluvial cones along the base of the ridge between Soldier Pass and the lake. Practically every one of these cones is so affected. Most interesting of all, however, is the area lying just southeast of the lake at the foot of the great scarp. The writer never before has seen such a wonderful exhibition of faulted alluvial cones. Any geologist interested in the phenomena of faulting would greatly enjoy a visit to this locality. The features of special interest are contained within an area $1\frac{1}{4}$ miles long and $\frac{1}{3}$ mile wide. Some of the principal faults of this area are indicated on the accompanying map. At the northern end of this area there is a fresh fault face 45 feet high at the very base of the mountain. About 30 feet of this scarp is sheer, with cemented fanglomerate forming the upper 10 feet of the cliff. This scarp no doubt resulted from very recent renewed movement along the Deep Spring fault.

A little south of the scarp last mentioned, an alluvial cone of

coarse débris is transected by a very distinct, straight fault trough about one-fourth of a mile long. The floor of this little fault depression is 100–125 feet wide, with a scarp reaching a height of 40 feet

FIG. 5.—View lengthwise through a fault trough (near center of picture) in an alluvial cone just east of Deep Spring Lake.

FIG. 6.—Freshwater pond in a fault trough in an alluvial cone east of Deep Spring Lake.

on the east side and 20 feet on the west side (Fig. 5). Several notches cut in the eastern scarp by drainage from the mountain side have small cones developed at their mouths in the fault trough.

A deep fresh-water pond several hundred feet long and fully 100 feet wide occupies part of a sharply defined basin formed by trough-faulting of an alluvial cone near the middle of the area here considered (Fig. 6). The scarp rises abruptly from the eastern shore of the pond to a height of about 30 feet, while the scarp on the west side of the pond is less abrupt and not so high. Slightly warm fresh-water springs feed the pond, which has an outlet through a small

FIG. 7.—A fault scarp 40 to 60 feet high in the large alluvial cone southeast of Deep Spring Lake

brook into Deep Spring Lake. This spring water no doubt rises from a considerable depth in the Deep Spring fault zone.

In the southern part of the little area under consideration, the largest cone near the lake ends abruptly in a nearly straight fault scarp about $\frac{1}{4}$ mile long, with a maximum height of over 50 feet. This impressive scarp, which is almost parallel with the Deep Spring fault, has been somewhat modified by erosion and by deposition of débris carried out of the canyon (Fig. 7). Some distance westward across a flat, small remnants of what was formerly the outer edge of this faulted cone are still visible.

Out in the mud flat east of Deep Spring Lake there is a long distinct sag several feet deep and several hundred yards long. This is of the nature of an incipient fault trough parallel to the Deep Spring fault.

DRAINAGE HISTORY

High-level, abandoned stream beds.—One of the most noteworthy geological features of the Deep Spring region is the occurrence of river boulders and gravels at several places on top of a desert mountain ridge. These river deposits are amazingly fresh and well pre-

FIG. 8.—River-bed boulders on top of the mountain, 1,000 feet above the floor of the valley, about 2 miles east-northeast of Deep Spring Ranch, at locality *w* on the map.

served. They occur on the ridge northeast of Deep Spring Ranch at localities *w*, *x*, *y*, and *z* indicated on the map. Very common among the materials are hard, fresh, well-rounded boulders of quartzite and rhyolite. The finest display is on top of the ridge at locality *w* at an altitude of 6,300 feet, or 1,000 feet above the valley floor (Fig. 8). Several acres are there covered densely with boulders and gravel, many of the boulders being a foot or more in diameter. A large, somewhat higher, area, at locality *x*, is more thinly covered with gravel and large pebbles. Some similar material is visible on the *col* at locality *y*, but apparently it is largely covered with débris which has either slid or washed down the adjacent steep slopes. The river deposits at locality *z* are highest of all. They lie at an altitude of 6,800–6,900 feet, or 1,500 feet above the floor of the adjacent valley.

This deposit includes many very smooth, hard pebbles and boulders up to a foot in diameter.

These high-level gravels more than likely mark former eastward courses of Wyman Creek before the displacement along Deep Spring fault took place to give rise to the present-day topography. Evidently the relative uplift of the ridge went on so rapidly that Wyman Creek was forced to flow across the rising mountain block farther and farther southward until finally the rate of uplift became so great that the creek was forced into the developing Deep Spring fault basin. The remarkable freshness of these river deposits, combined with their altitude, clearly indicates that most of the movement along the Deep Spring fault has taken place in Quaternary time.

Narrows of Wyman Canyon.—About 2¼ miles above the mouth of its canyon, Wyman Creek abruptly enters a narrows in solid rock. This gorge, 200 or more feet deep and only 15–20 feet wide at the bottom, has almost vertical walls. Down-cutting by the stream has here been very rapid on account of the crowding of the swift flood waters from the high mountains through this narrow defile. Evidently a small block, bounded sharply by a fault on the west (see map), has been uplifted across the well-established course of the stream without diverting it. In other words, it is a perfect example of an antecedent stream in this portion of its course.

Late Quaternary lake.—There is clear evidence that a fresh-water lake occupied Deep Spring Valley in relatively late Quaternary time. Remnants of the near-shore lake deposits occur at four rather widely separated localities, indicated by the letters *a*, *b*, *c*, and *d* on the map. These deposits are many acres in extent at *a* and *b*, where the character of the material may best be studied. The lake beds consist mainly of soft, nearly white, limy, stratified material which contains numerous well-preserved, tiny, gastropod and pelecypod shells, including a number of species. The deposits have been preserved to view only in those places which have been protected against either notable erosion or concealment under subsequently accumulated alluvial materials washed out of the mountains.

The lake beds range in altitude between 5,200 and 5,500 feet, with the latter figure almost, if not quite, reached at all four localities. It seems that the lake filled Deep Spring Valley to about the

5,500-foot contour line. The lake, therefore, had a length of 12½ miles; a maximum of 4½ miles; and a depth of probably not less than 400–500 feet. The fact that the lake beds all lie at practically the same altitude, almost undisturbed except by erosion, suggests strongly that the Deep Spring Basin had reached essentially its present-day condition at the time of the existence of the lake. It is true, however, that the very recent dislocation of the alluvial cones along the Deep Spring fault scarp has occurred since the disappearance of the lake.

The outlet of the Quaternary lake was no doubt across the mountain ridge through Soldier Pass at an altitude of about 5,500 feet, and into Eureka Valley. A fairly well-defined channel and also lake deposits exist in this pass, which is the lowest place on the rim of Deep Spring Valley.

Deep Spring Lake and springs.—Deep Spring Lake is a shallow body of water about a mile in diameter situated in the southern and lowest part of the valley. Its water is highly charged with various alkaline substances. Only at infrequent times of localized cloudbursts do surface waters flow over the desert floor of the valley and into the lake. The only perennial supply of water is the set of springs which emerge along the Deep Spring fault and minor associated faults east and south of the lake. These springs send several streams of slightly warm water into the lake.

EN ÉCHELON FOLDS AND ARCUATE MOUNTAINS

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ABSTRACT

En échelon folds and miniature arcuate mountain systems were produced experimentally in several ways. Differential stress transmission in the horizontal plane, through rigid materials bordering incompetent beds, gave rise to *en échelon* folds and arcuate systems even though non-rotational compression was applied. The same phenomena were also produced by applying differential (rotational) compression in the horizontal plane against homogeneous materials. Application of principles derived from the performed experiments are made to specific examples in nature.

INTRODUCTION

The problem of *en échelon* folds and the closely related phenomenon arcuate mountains is a complex but fascinating subject. Suggestions and theories as to their origin have already found their way into the literature of structural geology.[1] Some investigators have employed mathematical formulae in explaining the origin of *en échelon* folds, while others have called upon rotational stresses operative in the horizontal plane. What appear to be plausible suggestions as to the origin of certain types of *en échelon* folds and arcuate systems came to light during the experimental investigations recently completed by the writer in the geophysical laboratories of the University of Chicago.[2] It is not the purpose of this paper to discuss

[1] W. H. Hobbs, "Mechanics of Formation of Arcuate Mountains," *Jour. of Geol.*, Vol. XXII (1914), pp. 71–90, 166–88, 193–208. Warren J. Mead, "Mechanics of Geologic Structures," *Jour. of Geol.*, Vol. XXVIII (1920), p. 505. R. T. Chamberlin and F. P. Shepard, "Some Experiments in Folding," *Jour. of Geol.*, Vol. XXXI, No. 6 (1923), pp. 492–95. Lyndon L. Foley, "The Origin of the Faults in Creek and Osage Counties, Oklahoma," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. X, No. 3 (1926), pp. 293–300. M. King Hubbert, "The Direction of the Stresses Producing Given Geologic Strains," *Jour. of Geol.*, Vol. XXXVI, No. 1 (1928), pp. 75–84.

[2] Previous publications on the basis of these experiments are as follows: R. T. Chamberlin and T. A. Link, "The Theory of Laterally Spreading Batholiths," *Jour. of Geol.*, Vol. XXXV, No. 4 (1927). Theo. A. Link, "The Origin and Significance of Epi-anticlinal Faults as Revealed by Experiments," *Bull. A.A.P.G.*, Vol. XI, No. 8 (1927). Theo. A. Link, "The Relationship between Over- and Underthrusting as Revealed by Experiments," in the press, *Bull. A.A.P.G.* (1928).

previous contributions to this interesting subject, but to describe several ways in which *en échelon* folds and miniature arcuate systems were produced in the laboratory, to analyze them geometrically, and to point out where similar settings in nature may have produced analogous results. Each new idea is merely another stepping-stone toward the ultimate solution of this problem. General or widespread application is, in the writer's opinion, to be attempted with neither previous contributions nor with these offered in this paper. This was emphatically impressed upon the writer's mind during the performance of a series of some fifty experiments. In some of these, *en échelon* folds resulted frequently when no special attempt had been made to produce them, and in some cases the phenomenon was almost entirely lacking when attempts were made to produce such folds.

In the experiments performed two different methods to develop *en échelon* folds and arcuate systems were employed, namely, by applying stresses differentially in the horizontal plane against essentially uniform artificial sediments, or by applying non-rotational pressure in the horizontal plane against materials the consistency of which was not uniform, thus giving rise to differential transmission of stresses through the beds of different competency.

DIFFERENTIAL STRESS APPLICATION

Method A.—It is generally conceded that folds and related structures develop normal to the principal axis of stress. Therefore it should be possible to produce artificial folds in various alignments with respect to one another by varying the axes of applied stress. Figure 1 illustrates, in the horizontal plane, a push-block *A B C* whose face is, in part, the arc of a circle. If the sediments to be deformed are relatively incompetent, the first manifestation of pressure application is the development of a fold *X* in front of the straight face of the push-block. Upon further application of pressure the fold *X* will continue to grow toward the left, but with a tendency to parallel the curved face of the push-block. If the artificial sediments are slightly rigid the tendency to parallel the face of the push-block is diminished for obvious reasons. Continued pressure application puts into play stresses whose axes are not parallel

to those which produced fold X. At a given point on the curved part of the push-block the operative force may be resolved into component f which acts parallel to the direction in which the push-block is forced against the sediments, and component f' operative normal to the face of the push-block (Fig. 1). The actual operative force at this assumed point is F, the resultant of components f and f'. Fold X', developing opposite this given point, should lie normal to the principal axis of stress (line F–F). This is tangential to the curved face, and at an angle to the previously developed fold X. The extremities of fold X' tend to conform to the curved face of the push-block as shown in the figure. The resulting relationships between folds X and X' is what is generally known as an *en échelon* alignment. Such folds, or series of folds, may exhibit an *overlap* between their extremities or a *gap*. Since the maximum shortening was effected opposite B, more and larger folds should develop in front of B with a gradual diminution in size and number toward the left.

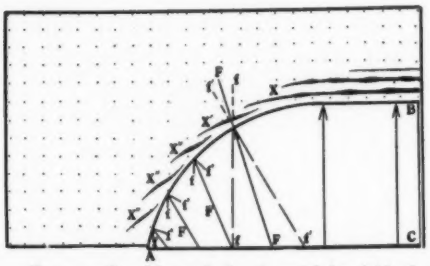

FIG. 1.—Graphic analysis of *en échelon* folds developed by forcing a push-block A B C against homogeneous material (dotted). Folds X develop normal to the straight front of the push-block. Fold X' develops normal to the axis of principal stress F–F which is the resultant of f–f and f'–f', since f–f is the direction in which the push-block is forced against the artificial sediments while component f'–f' acts normal to the curved face of the push-block. Folds X'' develop in the same off-set or *en échelon* relationship. The hooks on the arrows represent components f and f' as indicated.

Figure 2 illustrates, in the horizontal plane, the results of forcing against artificial sediments a push-block A B C whose face is the partial arc of a semicircle. The geometrical analysis is similar to that illustrated in the previous figure. An arcuate system of *en échelon* folds develops normal to the varying axes of stress (indicated by the arrows). In this arcuate system of folds there may also develop transverse tension faults or "flaws" (F) converging toward the locus of maximum compression.[1] Here again it is obvious that the

[1] Theo. A. Link, *op. cit.*, *Bull. A.A.P.G.*, Vol. XI, No. 8 (1927).

greatest number and the larger folds should develop in front of *B*, with a gradual diminution in size and number laterally toward *A* and *C*. An example of such an arcuate system of folds produced experimentally is illustrated in Figure 3. It is in place to suggest here that the sinuous trend of the Appalachian Mountains may owe its origin to a lateral variation in the tangential forces which gave rise to this system. The salients, so well described by Keith,[1] might indicate loci of greater compression while the "recesses" are the opposite. That compressive forces should not be uniform along the margins of the resulting uplifts requires little imagination. In fact, during the performance of these experiments it was found that once a locus of relief had been established, the tendency for a concentration of folding and faulting in that vicinity was greatly enhanced.

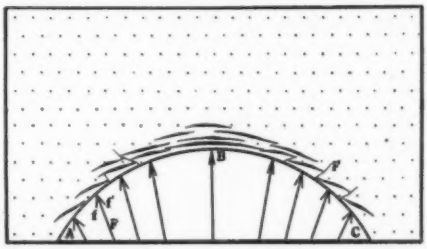

FIG. 2.—An arcuate system developed by forcing push-block *A B C* against artificial sediments (dotted). The folds tend to orient themselves normal to the operative forces at various points along the arc-shaped push-block. The resolution of forces is the same as explained in Figure 1. The hooks on the arrows represent components *f* and *f'*, while main line represents the resultant force *F*. Transverse flaws in the dotted area are indicated by *F*.

Method B.—Results similar to those just described were accomplished in a slightly different manner, as illustrated in Figures 4 and 5. On the side against which the pressure was applied the box bulged outwardly as indicated by *A B C* in Figure 4. The bulging part of the pressure box *A B C* was supported by a piece of strong flexible leather. Against this was forced a straight board or push-block, *P*–*P*. Obviously in this case the first manifestation of pressure application is the growth of folds in front of *B*. Although the side of the pressure box is convex, nevertheless the pressure is applied normal to the face of the push-block, as indicated by the arrows in Figure 4. Because of the leather side moving outwardly toward the right and

[1] Arthur Keith, "Outlines of Appalachian Structure," *Bull. Geol. Soc. Amer.*, Vol. XXXIV (1923), pp. 309–80.

left, as indicated by the longer arrow, a lateral component is introduced and the actual operative forces radiate from a point below *B* as indicated by the dashed arrows. The results of such forces are manifested clearly in the model by the line of folds extending far toward the upper left-hand corner and almost equally as distant on the right side, as well as the transverse "flaws" (*F*) converging toward the locus of maximum compression. This minia-

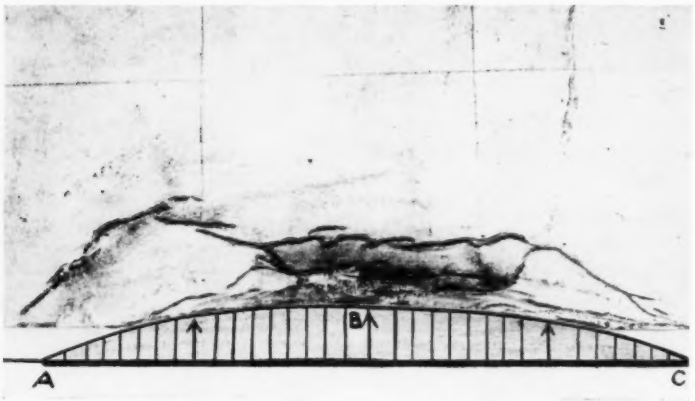

FIG. 3.—Top view of Experiment No. 19. A miniature arcuate system produced experimentally by forcing a push-block *A B C* (shaded), whose face is curved, against artificial sediments. The direction of movement of the push-block is indicated by the arrows. The resolutions of forces is as explained in Figures 1 and 2. The squares represent 10 cm.

ture system differs from the one illustrated in Figure 3 in that only the outer margin is arcuate while the margin in front of the push-block parallels the latter. Without the knowledge of the method employed in producing such an uplift an observer might find difficulty in explaining its origin; however, the transverse flaws converging toward the locus of maximum compression, the outer margin presenting its concave side toward the push-block, and the cross-section (Fig. 5) exhibiting the low-angle overthrusts, is sufficient evidence to conclude that forces dying out laterally gave rise to this miniature system.[1] The writer knows of no case in nature

[1] Theo. A. Link, *op. cit.*, *Bull. A.A.P.G.*, Vol. XI, No. 8 (1927).

which might have had an origin similar to the artificial system just described. It is merely presented here to illustrate the various ways in which arcuate systems may be developed.

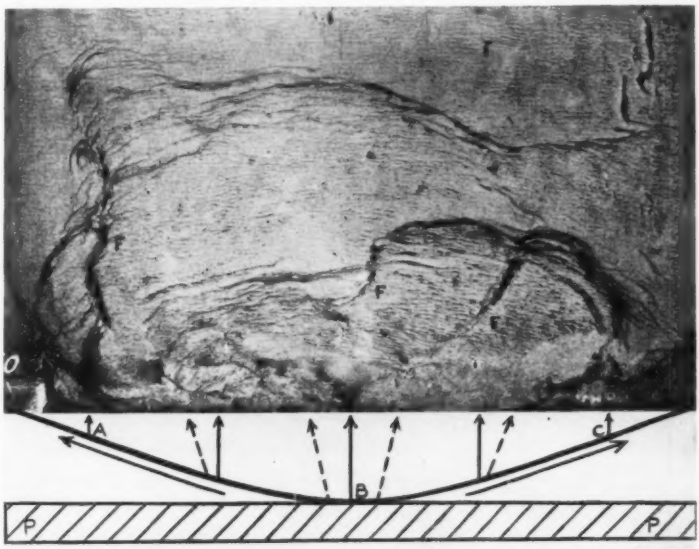

FIG. 4.—Top view of Experiment No. 50. A miniature arcuate system produced experimentally as indicated by the arrows. Before application of pressure the box containing the artificial sediments bulged outwardly along the side *A B C* which was made of strong leather. *P–P* was then forced against the leather side until the bulge was taken out. Note in particular the outer arcuate margin and the transverse "flaws" converging toward the center of the lower side where the maximum shortening was effected. Size of model 13×22 cm.

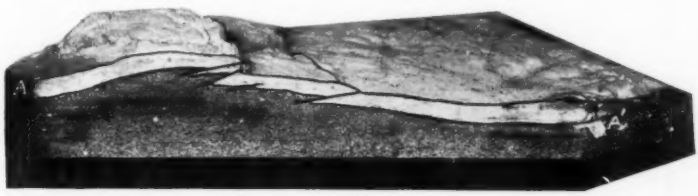

FIG. 5.—Upper part of cross-section cut through model illustrated in Figure 4. Note the low-angle overthrust faults.

DIFFERENTIAL STRESS TRANSMISSION DUE TO LATERAL VARIATION OF SEDIMENTS

En échelon folds resulted during the performance of many experiments when no special attempt was made to cause their development. An example of such a case is illustrated in Figure 6. Evidently unintentional variations in the rigidity of the artificial strata near or at the surface, causing them to transmit stresses irregularly, played an important part in the orientation of the folds. Upon examination it was found that the uppermost plaster of Paris layers had faulted at various angles in the horizontal plane, and the over-

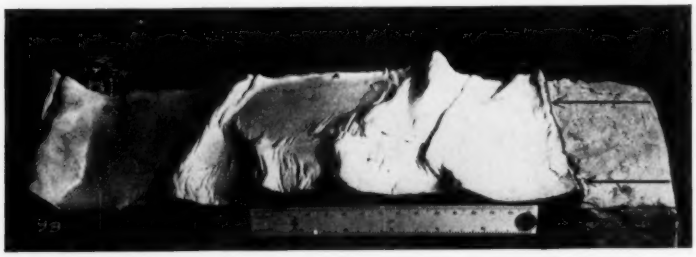

FIG. 6.—Top view, taken at an angle, illustrating *en échelon* folds which resulted in Experiment 9B without any attempt to cause their development. Pressure was applied from the right as indicated by the arrows. Unequal rigidity in the artificial sediments evidently gave rise to differential stress transmission. The scale represents 15 cm.

lying, less rigid, damp blotting paper layer developed folds above these fault planes in typical *en échelon* alignment. To test the validity of these conclusions a series of experiments were subsequently performed in which the artificial sediments were prepared with known lateral variations in their rigidity or stress-transmission ability. The lateral variation of sediments in nature is an established principle of geology and needs no demonstration here to justify its application to structural problems. It is the rule rather than the exception.

Artificial sediments which are of such consistency that they are unable to transmit stresses any considerable distance invariably are folded or faulted in the area immediately in front of the push-block. This was established years ago by Bailey Willis in his classic experiments on "The Mechanics of Appalachian Structure."[1] It is obvious

[1] Bailey Willis, *13th Ann. Report, U.S. Geol. Survey*, Part II (1893), pp. 247.

that if certain areas of the artificial strata (considered in the horizontal plane) are of such consistency that stresses are transmitted for some distance, there will result an irregular line of folds in front of the push-block. In the weaker sediments the folds and faults will develop directly in front of the push-block, while the more rigid will remain undisturbed and transmit stresses further on, developing folds where weaker material is first encountered. In this manner *rotational* stresses are put into play in the horizontal plane resulting in belts of *en échelon* folds or arcuate systems. This may happen even though purely *non-rotational* compression is *applied* in the horizontal plane.

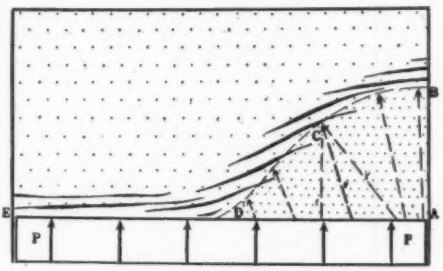

FIG. 7.—Illustrating in the horizontal plane how inequality in the rigidity of materials may give rise to *en échelon* folds by differential transmission of stresses. The closely dotted part of the layer occupying area *A B C D* is more rigid or competent than the remaining portion. Application of pressure, as indicated by the heavy short arrows in the push-block *P–P*, causes a transmission of stresses (dashed arrows) to the edge of the more rigid mass *B C D*, where folds grow simultaneously with those developing immediately in front of the push-block between *D* and *E* in the less rigid portion of the stratum.

These principles are best illustrated by diagram in Figure 7. Push-block *P–P* is forced against the artificial strata as indicated by the heavy arrows. The closely dotted area *A B C D* is occupied by relatively rigid material, while the remainder is composed of relatively incompetent material. Upon application of pressure, stresses are transmitted through the rigid material to its margin *B C D*, as indicated by the longer arrows. Deformation of the less competent beds results along *B C D*, while the equally incompetent beds are simultaneously deformed directly in front of the push-block between *D* and *E*. The resolution of forces along the face *B C D* is essentially the same as explained in Figure 1, since the conditions herewith described amount to nothing more than having a push-block whose face is *B C D E*. Therefore, a belt of *en échelon* folds should develop along *B C D* very much in the same manner as explained in Figure 1.

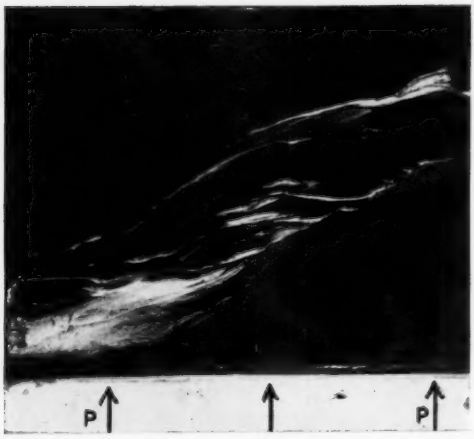

FIG. 8.—Top view of Experiment No. 45, illustrating a belt of *en échelon* folds produced by applying non-rotational pressure, as indicated by the arrows, against artificial sediments with a lateral variation in their rigidity. In the lower right-hand corner of the box the underlying blue clay layer was considerably more rigid than in the remaining part of that layer. Model measures 15×22 cm.

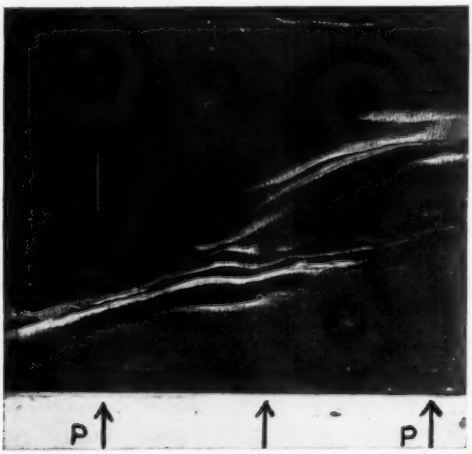

FIG. 9.—Top view of Experiment No. 46, illustrating another belt of *en échelon* folds produced with essentially the same sedimentary setting and the same application of non-rotational pressure as in Figures 7 and 8. Model measures 15×22 cm.

Several experiments, in which the previously described setting was used, were performed. An inch layer of semiplastic blue clay, mixed with cement, was placed in a pressure box. In the right-hand corner of the box, covering an area corresponding roughly to *A B C D* of Figure 7, sufficient cement was added to stiffen this portion of the blue clay considerably. Over all of the blue clay was

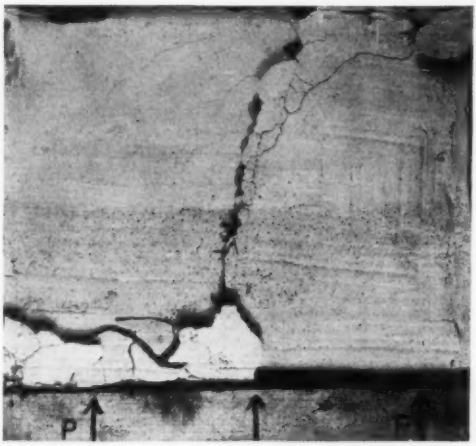

FIG. 10.—Top view of Experiment No. 44 after removal of the overlying tissue paper. In this case the rigidity of the blue clay layer in the right portion of the box was great enough to transmit stresses to the more plastic parts on the opposite side, thus giving rise to the horizontal fault, above which *en échelon* folds developed in the tissue paper. On the left side the blue clay was less rigid and buckled up directly in front of the push-block. Model measures 15×22 cm.

spread a quarter-inch layer of incompetent material consisting of a mixture of molten paraffin and grease (petrolatum), which was then allowed to solidify. A thin piece of crêpe or tissue paper was then placed on the surface of the paraffin-grease layer. An overburden of loose, dry, fine-grained sand was deposited over all of this to a depth of about two inches. *Non-rotational* compression was then applied as indicated by the arrows in Figures 8 and 9, and the belts of *en échelon* folds illustrated in these figures resulted. In one of the experiments the blue clay was of such rigidity as to form a horizontal fault, above which developed *en échelon* folds in the tissue paper

(see Fig. 10).[1] Belts of *en échelon* folds very similar to those described and pictured herewith are a common occurrence in nature. They may be found along the outer margins of nearly all orogenic units. They are especially common in the Foothills Belt east of the Rocky Mountains in Alberta and Montana. The presence of deltaic deposits, embayments, or promontories which give rise to rapid

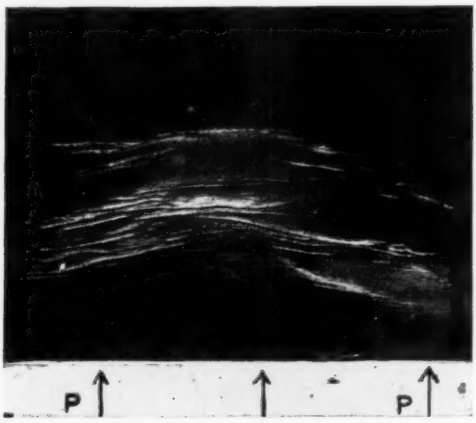

FIG. 11.—Top view of Experiment No. 49, illustrating a miniature arcuate system produced by applying non-rotational pressure, as indicated by the arrows. In the center, directly in front of the push-block in the shape of a small arc, the underlying blue clay layer was stiffened considerably, thus causing a greater transmission of stresses centrally dying out laterally to the right and left. Compare this artificial system with that of the Jura Mountains illustrated in Figure 12. Model measures 15×22 cm.

lateral variations of sediments along the ancient shore lines afford ideal settings for the development of *en échelon* folds in a manner analogous to those just described.

Arcuate systems of folds were developed experimentally in a similar manner with the same artificial sedimentary setting save that in the lower clay layer the more rigid portion had the outline of an arc or delta similar to the push-block illustrated in Figure 2. The result of such an experiment is illustrated in Figure 11. A comparison of the artificially produced arcuate system in Figure 11 with

[1] L. S. Foley, *op. cit.*, *Bull. A.A.P.G.*, Vol. X, No. 3 (1926).

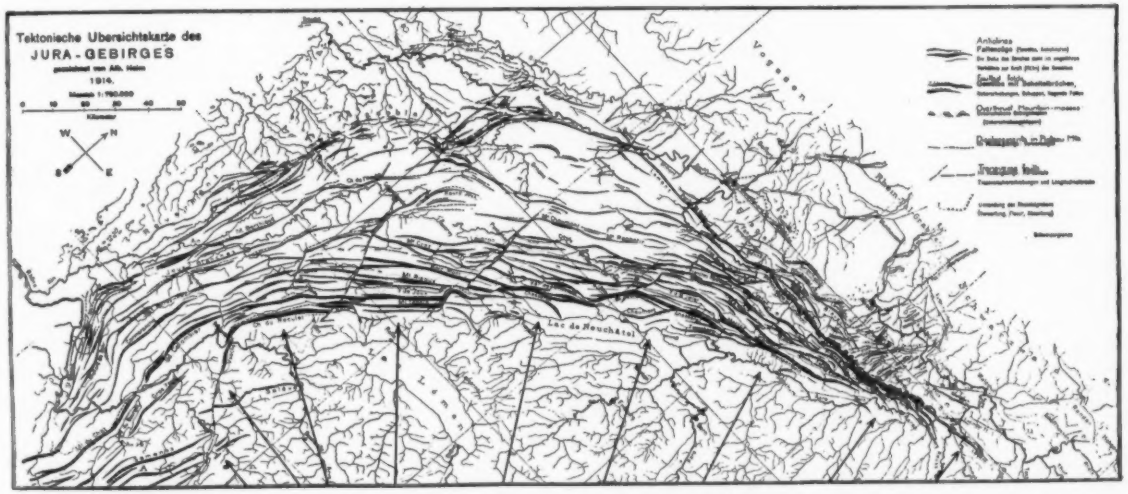

FIG. 12.—Tectonic map of the Jura Mountains of Europe taken from Heim's *Geologie der Schweiz*. The arcuate shape of this system, the *en échelon* alignment of folds, and the transverse "flaws" are conspicuous. The probable direction of stress transmission is indicated by the arrows. Compare this map with the miniature arcuate system produced experimentally and illustrated in the previous figure.

that of the Jura Mountain Arc (Fig. 12) brings out a striking similarity, and in consideration of the actual sedimentary setting giving rise to the Jura Mountains the analogy is even more complete. Between the Alps and the Jura Mountains is a valley underlain by a thick, rigid, and competent bed of conglomerate (Nägelfluh) which is relatively undisturbed. During the late stages of the building of the Alps stresses were transmitted through the competent Nägelfluh conglomerate to the relatively incompetent beds in which folding and faulting took place giving rise to the Jura Mountains, an arcuate system bordering the outer edge of the Nägelfluh conglomerate.

SUMMARY AND CONCLUSIONS

The results of the experiments herewith described indicate that:

1. *En échelon* folds and arcuate mountain systems are polygenetic.

2. They may be due to a lateral variation of tangential forces, which amounts to rotational compression in the horizontal plane. The manner in which this may be accomplished is also variable.

3. They may be caused by non-rotational forces operative in the horizontal plane against sediments which have a decided lateral variation in plasticity, rigidity, or competency, giving rise to variable stress-transmission.

MELIKARIA: VEIN COMPLEXES RESEMBLING SEPTARIA VEINS IN FORM

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ABSTRACT

Melikaria are chemical deposits of honeycomb form typically found in the Quaternary of Brazos County, Texas. They differ from septaria in several ways. Chief among these is their development directly in the surrounding formation, without the medium of concretion development. The ground-water conditions and the existence of voids for the reception of the deposited silica are favorable to the continuance of their development to the present time. These deposits may serve as valuable guides in the interpretation of past and present ground-water and sedimentary conditions.

INTRODUCTION

The early Quaternary alluvial deposits of the uplands of Brazos County, Texas, are rich in various types of concretions and allied structures. Among these are many small, septaria-like, silicious vein-plexuses of honeycomb form (Fig. 1).

These objects were at first believed to be the veins of eroded septaria, but detailed field and microscopic study has proved them to be distinct structures. It becomes necessary, therefore, either: (1) to designate them as "septaria-skeletons," or (2) to extend the term "septarium" to include both: "Concretion possessing veins or filled cracks,"[1] and the dissociated veins and filled cracks, or (3) to coin a new term to apply to them. As they clearly are neither remnants of septaria, nor septaria in the early process of formation, the term "septaria-skeletons" would be misleading. The term "septaria" is so well established as applying to: "Concretions the interiors of which have parted, and the gaping cracks become filled with calcite or other mineral deposited from solution,"[2] that any broadening and redefining of the term would not be justified. The new term "melikaria"[3] is, therefore, proposed for them.

[1] W. A. Tarr, in *Treatise on Sedimentation*, by Twenhofel *et al.* (1926), p. 502.

[2] Chamberlin and Salisbury, *Geology*, Vol. I (2d ed. revised, 1904), p. 473.

[3] From the Greek, μελί-κηρον, "honeycomb." Adopted by the writer after consultation with D. B. Cofer, of the English Department, A. & M. College of Texas.

The melikaria described in this article differ from septaria in the following ways: (1) they are not associated with concretions, (2) they are of quartz, which, while not unknown as a septaria mineral, is far less common than calcite, and (3) the crystals have not grown inward perpendicularly to the wall, as is common in the crystals of septaria (Fig. 2).

FIELD AND LABORATORY DATA

Melikaria have been collected from several places in Brazos County, but the majority of specimens have come from a single bed, about three feet thick, exposed along the sides of an unnamed creek 2 miles west of College Station. The bed lies close to the bottom of the Quaternary deposits, and on the average less than 5 feet above the contact with the Yegua (Eocene) formation, the upper part of which contains many dense, impervious beds of clay. The average depth of the melikaria-bearing bed below the surface is about 10 feet.

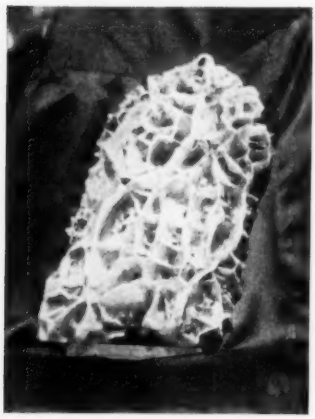

FIG. 1.—One of the largest specimens, 15×8×4½ inches.

The melikaria are all flattened structures of honeycomb form, and lie with their greatest dimensions in the horizontal plane. They vary considerably in shape, the commonest forms being round, oval, roughly heart-shaped, or scalloped. The symmetry is in general well marked, irregular forms being rarely found. The horizontal dimensions vary from 5 to 21 inches, and in all recorded cases the vertical dimension is between 1 and 4½ inches. When taken from the ground, the cells of the honeycomb are packed with sand and clay from the containing formation.

The walls of the honeycomb vary from $\frac{1}{16}$ to $\frac{1}{2}$ inch in thickness and are composed of dense, milky-white quartz. So compact are these walls that their external appearance is similar to that of novaculite. On fracturing, the interiors present a very fine, drusy structure. Geode cavities up to $\frac{1}{4}$ inch across occur in the walls with

some degree of frequency. The microscopic structure of these walls is illustrated in Figure 2.

The horizontal distribution of these forms is most striking and significant. They have been found along a few very narrow belts only. These belts are not known to be more than 50 to 100 yards across. Between these belts, or lines as they are hereafter designated, melikaria are not known to exist. In fact a careful examination of

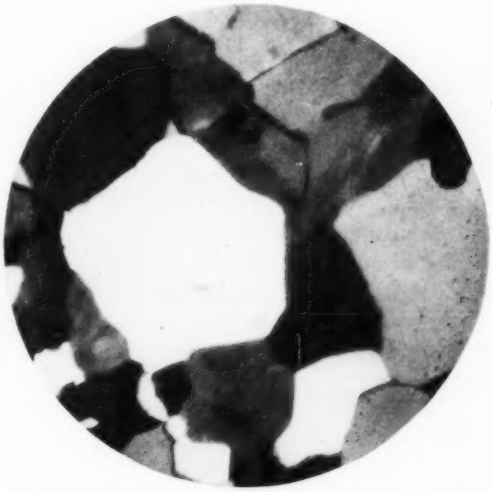

FIG. 2.—Thin section of wall as seen under low power with crossed nicols (80×). Quartz crystals variously oriented, and smaller toward the outside (left) of the wall. Some of the crystals show minute inclusions.

the lower Quaternary exposures indicates that they do not exist between them. Everywhere below these lines of melikaria development, fracture zones are found to be developed in the underlying impervious beds of the Yegua. These fracture zones are evidenced by jointing, by the existence of springs, and by their influence on drainage features. Due to its Quaternary covering, the trace of the fracture pattern of the Yegua can be neither accurately nor completely platted, but so far as it is known the pattern of melikaria development appears to be directly superimposed upon it.

Associated with the melikaria is much petrified wood, selenite,

and limonite, and traces of halite. The selenite and halite are found on the walls of the creek gorges as efflorescent mineral matter. The selenite is clearly secondary in the formation, and, together with the halite, has been deposited by deep waters as they have risen to the surface. The limonite occurs as concretionary material, as inclusions in the efflorescent minerals, and as secondary coatings on the clastic grains of the formation. In all these cases it clearly represents a chemical deposit from rising waters. The petrified wood is composed of three silica minerals: quartz, chalcedony, and opal. The water of the underlying formations is at present alkaline in reaction and is bearing silica in solution, probably in the form of sodium silicate.[1] The melikaria lie above the summer level of the water-table and over fracture zones along which the rising waters responsible for these various chemically deposited minerals are concentrated. In this position they occupy the place where the deep waters approaching the surface in the dry summer season begin, through evaporation, to deposit their less soluble constituents. It is at approximately this same level that rising silica-burdened, alkaline waters in the wet, winter season first come in contact with the soil-water charged with carbon dioxide and the acids of plant decay.

THE ORIGIN OF THE MELIKARIA

The problem of the origin of these forms involves two alternatives: (1) their erosion from cracked concretions and transportation to their present position, and (2) their formation *in situ*. Suggestive of the former must be mentioned: (1) their localized areal distribution along narrow belts, (2) their horizontal arrangement within the beds, and (3) the presence of "turtle stones" in the weathered Eocene formations to the north and west. Indicative of their formation *in situ* must be cited: (1) their apparently uneroded condition, (2) their usually unbroken symmetry, (3) the fact that the known "turtle stones" would not, by erosion, yield a form of septaria-skeleton like a honeycomb, (4) their close association with petrified wood, showing the former circulation of silica-bearing waters, (5) the consistency with which their horizon approximately coincides with the summer

[1] H. R. Brayton, professor of chemistry at A. & M. College of Texas. Verbal communication to the writer, September 26, 1927.

water-table level and the winter line of separation between the acidified soil-water and the deeper alkaline circulation, (6) their limitation to lines above fracture zones, and (7) the fact that during the hot, dry, summer months the ground cracks to a depth greater than that at which they are found. The cracks thus formed average about the width of the thickness of the melikaria walls.

From the foregoing facts the following conclusions concerning their origin seem warranted: (1) they were formed in place as epigenetic structures during late Pleistocene and Recent times. (2) They developed in the bottoms of deep, desiccation cracks, (*a*) by the acidification of silica-bearing waters rising from below through fractures during the wet season, and (*b*) by precipitation through evaporation above the water-table from deep, rising waters during the dry season. (3) It is very probable that their formation still is going on.

APPLICATIONS TO STRUCTURAL AND SEDIMENTATION PROBLEMS

In the light of the foregoing discussion we may define melikaria as small networks of veins of honeycomb appearance, consisting of mineral matter deposited from rising, aqueous solutions, and produced in the inclosing formation without the medium of concretion development. Although structures of this type are not known to the writer to have been previously described, the conditions necessary for their development are theoretically so widespread that their bearing on the interpretation of structural and sedimentation problems should be worth considering.

From their mode of origin it appears they must be limited to areas which were semi-arid, or had a long, annual, dry season alternating with a wet season, at the time of their development. In such areas they could not have developed at a depth greater than that of the level of the dry season water-table. On the other hand they could not have been developed at levels greatly above this, except along lines of vigorous fissure or tubular springs. If the approximate date of their development can be determined, they serve as a climatic indicator for that particular date and region. If melikaria are found composed of gypsum, barite, or other readily cold-soluble minerals, they must indicate a continuation of arid or semi-arid conditions

from the time of their development to the present. Silica melikaria, on the other hand, may remain, once formed, through subsequent climatic changes during which other climatic indices might be completely removed.

Their broadcast distribution within a formation could occur only if the underlying formations were porous conglomerates or sandstones. A development restricted to lineal zones may mark lines of solution channels if the underlying formation is a limestone or other soluble rock. In the case discussed in this paper, where the underlying formation is insoluble and has extensive impervious horizons, their arrangement in well-defined lines must mark master joints, or shear or fault zones below.

Finally, their mineralogical character and their vertical relations to the cementing, concretionary, and efflorescent minerals of the inclosing formation serve as a key to the groundwater conditions at the time of their development. This in turn may throw some light on the possible character of the deeper rocks of the area. No geologist experienced in field interpretation, however, needs to be warned against pushing these deductions too far, but as circumstantial and complementary evidence they may be valuable when conservatively used.

PRODUCTION OF TRACHYTE AND PHONOLITE FROM PYROXENE ANDESITIC MAGMA ASSOCIATED WITH LIMESTONE

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ABSTRACT

Most of the Javanese volcanoes have produced pyroxene andesites and basalts only. In the Volcano Merapi, which is a typical representative of this kind, trachyte, phonolite, and phanerites with preponderant leucite have been produced in association with a large xenolith of metamorphosed limestone. Similar reactions on a larger scale may have contributed to the production of the alkaline rocks of the East Indian Archipelago.

The belief that alkaline magma may be produced from basaltic magma by reaction with limestone has been put forward by Daly.[1] The formation of feldspathoids is considered due to a desilication of the magma, because the heavy lime-silicates will tend to carry silica downward. It is thought that escaping carbonic acid and other resurgent gases cause upward movement of the alkalies, and in this way alkaline magmas would be produced as differentiates from subalkaline magmas.

Bowen believes[2] that differentiation is controlled by crystallization. The sinking of crystals and the squeezing out of residual liquid are considered the all-important instruments of differentiation. Under the action of these processes typical igneous rock series would be formed from basaltic magma if it crystallized slowly enough. Limestones and alkaline rocks are often intimately associated, but according to him there is no assurance that the magma was not already an alkaline magma before it acquired this association.

Daly and Bowen agree that peridotite and augite andesite can

[1] R. A. Daly, "Origin of Alkaline Rocks," *Bull. Geol. Soc. America*, Vol. XXI (1910), p. 87; "Igneous Rocks and Their Origin," *ibid.* (1914), p. 410; "Genesis of Alkaline Rocks," *Jour. Geology*, Vol. XXVI (1918), p. 97.

[2] N. L. Bowen, "The Later Stages of the Evolution in Igneous Rocks," *Jour. Geol.*, Vol. XXIII, supplement No. 8 (1915); "Crystallization-Differentiation in Igneous Magmas," *Jour. Geol.*, Vol. XXVII (1919), p. 393; "The Behavior of Inclusions in Igneous Magmas, *Jour. Geol.*, Vol. XXX (1922), supplement, p. 513.

be formed through gravitative differentiation from a primary basaltic magma. For the formation of more remote derivatives, however, their conceptions diverge. Daly emphasizes the assimilation of foreign material, while Bowen's synthesis furnishes us with a magma rich in alkalies without assimilation. Bowen does not deny that some alkaline rocks may be formed by solution of carbonates.

In the present state of the problem direct tests by experiments would be desirable. Niggli's[1] diagrams of the system $K_2O-CaO-SiO_2-CO_2$ throw some light on the reactions, which result from the assimilation of limestone in silicate magmas. To approximate the conditions of nature in experiments on the effects of dissolving limestone in basaltic melts will be difficult. Therefore I gave special attention to the effects of basaltic and andesitic magma on inclusions of limestone when studying the xenoliths of the volcanoes of the East Indian archipelago.

The effects of volcanic magma on xenoliths approach those of experiments under the conditions of nature, and it is of interest that trachyte and phonolite can be formed from pyroxene-andesitic magma in association with limestone. Many inclusions have been found in the rocks of the volcano Merapi (Java), particularly in those of the lava dome, which forms the top of the cone. The rocks of the volcano are pyroxene andesites; they often have a small olivine content; and in some specimens amphibole is present. In many xenoliths which were studied under the microscope the common minerals wollastonite, diopside, garnet, anorthite, and sometimes epidote were found. Calcite is also present in some of these metamorphic inclusions.

A large block ($50\times50\times30$) centimeters of metamorphosed limestone on the slope of the volcano Merapi has a zonal structure. In some zones wollastonite is the principal mineral; calcite and augite are often found associated with it, and sometimes plagioclase and augite are the predominating minerals. In many zones a colorless mineral which is nearly isotropic or which shows distinct optical anomalies is found in varying quantities. Microchemical tests and

[1] P. Niggli, Gleichgewichte zwischen TiO_2 und CO_2, sowie SiO_2 und Co_2 in Alkali-, Kalk-Alkali-, und Alkali-Aluminatschemelzen, *Zeitschr. f. anorg. u. allgem. Chemie* (1916), p. 241.

the determination of the refractive index prove that much of it is leucite. Sometimes biotite is found associated with it, together with augite and plagioclase. Zones consisting of orthoclase and augite with some calcite are also found. Among the leucite-bearing zones the extremely scarce phanerites with preponderant leucite are represented.

Of particular interest are zones of porous trachyte and leucite phonolite with much glass in the groundmass. Two main types can be distinguished: (1) Phenocrysts of orthoclase are imbedded in a groundmass of orthoclase or acid plagioclase, augite (partly aegirine-augite), and glass. (2) The groundmass contains the same constituents as that of 1, but the principal mineral is leucite in partly idiomorphic or rounded crystals with small inclusions. They give a strong potash reaction and show positive double refraction with polysynthetic twins. Sometimes leucite forms small phenocrysts, but those of orthoclase are absent. In both types calcite, apparently of magmatic origin, is found and they both pass into mineral associations which are found in other zones of the metamorphosed limestone.

Of course these observations have not the same value as that of an experiment on the effects of dissolving limestone in basaltic or andesitic melts. We have not directly observed that the alkaline magma was not already the same alkaline magma before it was associated with the limestone. But if trachytic and phonolitic magma originated in this volcano of Java without reaction with limestone, we would expect that at least a single piece of alkaline rock would have been found among the products of the numerous large volcanoes from which pyroxene andesites and basalts have been ejected. These expectations do not agree with the facts.

The parts of the magma which are found associated with limestone seem to have been very fluid and rich in fugitive constituents, and alkalies may already have concentrated. This is not astonishing where a magma reacts with its wall rocks and causes metamorphism under the influence of magmatic emanations. Large quantities of alkaline magma could develop by accumulation of fugitive constituents and reaction with highly calcareous sediments because an important addition of lime will not be prevented then by the

progressive crystallization of the magma.[1] As on a minor scale in the case of the volcano Merapi, new phases might appear in sufficient quantity during the long periods of dormancy of a volcano. Such reactions may have contributed to the production of the alkaline rocks of Java. The prevailing rocks of the numerous volcanoes of this island are pyroxene andesites and basalts. The rocks of the volcano Merapi in Central Java belong to this prevailing type. Alkaline rocks are restricted to the neighborhood of the north coast of the eastern part of the island (Muriah near Semarang, island of Bawean to the north of Surabaia, Lurus to the west, and Ringgit to the east, of Besuki). Subalkaline rocks are found among the products of the same volcanoes, and the syngenesis of subalkaline and alkaline rocks is particularly emphasized by the local occurrence of trachyte and phonolite among the products of the volcano Merapi. Our knowledge of the substratum is incomplete, but for some of the alkaline rocks there certainly has been a great possibility for reaction with limestone. A generalization of the observed facts seems to be premature. The course of differentiation may have been influenced by other causes. Smyth[2] stated in his paper on the origin of alkaline rocks that their differentiation has been largely effected through the agency of mineralizers and that the several hypotheses have certain features in common, despite great differences of emphasis.

The main point is that most of the Javanese volcanoes have produced pyroxene andesites and basalts only, and that in a typical representative of this kind trachyte and phonolite have been produced in connection with limestone.

[1] S. J. Shand, "The Problem of the Alkaline Rocks," *Proc. Geol. Soc. South Africa*, (1922), p. xxx.

[2] C. H. Smyth, "The Chemical Composition of the Alkaline Rocks and Its Significance as to Their Origin," *Amer. Jour. Science*, Vol. XXXVI (1913), p. 45.

THE PREPARATION OF THIN SECTIONS OF FRIABLE ROCK

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University of Chicago

ABSTRACT

The impregnation of friable rocks with bakelite varnish, followed by a period of baking in an oven, produces a very coherent rock-chip from which thin sections can be readily made. This method has advantages over others in that the resulting rock-chip is very hard and tough and is not affected by heat. The detailed technique is described.

INTRODUCTION

In attempting a petrographic study of the Pleistocene deposits, the writer was confronted with the problem of preparing thin sections of some very incoherent materials, many of them so friable that they are reduced to a powder by lightly brushing with a camel-hair brush. Johannsen[1] has cited and briefly described more than a dozen procedures used by earlier workers in preparing thin sections of friable materials. These older methods, as well as the newer ones described below, all have certain objectionable features which are now rather easily overcome.

Schlossmacher[2] impregnated porous rocks in vacuum by placing the specimen on a shoulder of a vessel, the bottom of which contained Canada balsam. After the air had been removed from the pores of the rock, the vessel was tipped, allowing the rock-chip to drop into the balsam.

In 1921, Sayles[3] described a method of preparing thin sections of till and stratified clay. The specimen was soaked in chloroform in vacuum for several hours and allowed to stand over night at 50° C.

[1] Albert Johannsen, *Manual of Petrographic Methods*, pp. 599–602. McGraw-Hill Book Co., 1918.

[2] K. Schlossmacher, "Ein Verfahren zur Herrichtung von schiefrigen und lockeren Gesteinen zum Dünnschleifen," *Centralblatt für Mineralogie, Geologie und Paläontologie* (1919), pp. 190–92.

[3] Robert W. Sayles, "Microscopic Sections of Till and Stratified Clay," *Bull. Geol. Soc. Amer.*, Vol. XXXII (1921), pp. 59–62.

Powdered Canada balsam was gradually added for a period of a week. The chloroform was then driven off by heating on a water bath for a day or two, and finally the balsam was cooked on a glycerine bath. The method is rather involved and requires over ten days to complete the procedure. The final section is not as firm as the ordinary rock-specimen and considerable care must be taken in grinding or disruption will occur.

Ross[1] described a method of impregnating porous rocks with bakelite varnish. He soaked the specimen in the varnish which had been diluted with ether and methyl alcohol (half and half). After good penetration had been secured, the rock-chip was allowed to stand for a day or two, permitting the solvents to evaporate, and finally baked in an oven for two days at a temperature of 70° to 100° C., causing the varnish to harden. The very tough chip was then ground thin with an abrasive and kerosene, or nujol, in place of water. According to Ross, "No method of impregnating porous material results in such great coherence, toughness, and hardness as the bakelite treatment."

The method about to be described is somewhat similar to that of Ross except for the following: (*a*) Methyl alcohol is used as a solvent instead of ether and methyl alcohol. This results in more rapid and complete impregnation. (*b*) By not using ether as a solvent, the soaking can be carried on under low pressure. This also increases the rapidity of penetration. (*c*) Evaporation of the solvent takes place under reduced pressure, thus eliminating the necessity of allowing the specimen to stand for a day or two so that the solvent may pass off. (*d*) The increasing viscosity of the varnish resulting from the evaporation of the solvent under low pressure causes the pores of the outer shell of the rock to be filled with a more concentrated varnish, which, after baking, leaves the interstitial openings almost completely filled with the hardened bakelite. The advantages of this feature will be discussed further in a later part of the paper. (*e*) More complete impregnation resulting from (*a*), (*b*), and (*d*) allows the use of water in grinding many of the colloidal clays. (*f*) By using benzene instead of kerosene or medicinal oil in

[1] Clarence S. Ross, "A Method of Preparing Thin Sections of Friable Rock," *Amer. Jour. Sci.*, 5th Ser., Vol. VII (1924), pp. 483–85.

grinding down the flat face which is to be cemented to the glass slide, it is easier to obtain a firm cementation.

Although no definite procedure can be applied to all friable materials, it is believed that workers using the bakelite method for the first time will find it easier to modify an already developed procedure to fit their needs than to develop their own technique. For this reason the procedure is described in greater detail than that given by Ross.

Ross[1] has also described a method of impregnating porous rocks using kollolith in place of bakelite. Kollolith is a preparation resembling Canada balsam, although it is tougher, does not bubble when heated, and may be kept at a temperature of 170 C. for a considerable length of time without darkening. The technique involved is similar to that used in the bakelite method, although it is not so time-consuming. The resulting rock-chip is not so tough and coherent as one produced by impregnating with bakelite.

It has been pointed out by R. E. Head[2] that the bakelite method of impregnating porous rocks has proved to have an application in the polished section field as well as in that of the preparation of thin sections. The method described below may be easily adapted to the preparation of polished sections of rocks of a highly fractured nature. Head[3] has also described a method of preparing polished sections of mill concentrates, middlings, etc., by the use of redmanol, a bakelite product. The briquets produced by this method have advantages over those made with sealing-wax in that a much higher polish may be secured and that the binder is resistant to acids and other reagents commonly used to identify minerals in the polished condition.

THE BAKELITE METHOD

Bakelite is a synthetic phenol resin product manufactured by the Bakelite Corporation, Chicago, Ill., to whom acknowledgment is here made for much helpful advice and co-operation. The product

[1] Clarence S. Ross, "Methods of Preparation of Sedimentary Materials for Study," *Econ. Geol.*, Vol. XXI (1926), pp. 460–68.

[2] Personal communication.

[3] R. E. Head, "A New Method of Preparing Briquetted Mineral Grains for Microscopic Study," *Engineering and Mining Journal-Press*, May 30 (1925), pp. 889–90.

is sold in the form of a varnish at a cost of about 45 cents per pound. Upon applying heat, the varnish polymerizes, forming a new compound which is hard, tough, resistant to fracture, and not affected by heat. In thin section, bakelite appears as a pale yellow isotropic substance of high relief. Its high index of refraction (1.60–1.64, varying with the degree of polymerization and the freshness of the varnish) is in general an objectionable feature, although it accentuates the pore spaces and reduces the relief of minerals of high index. The varnish should be kept in a cool place and should not be stored much over three months. It becomes darker, heavier, and more viscous with age.

The specimen to be sectioned, if not too friable, is ground down to a flat surface on dry emery cloth. The loose dust remaining on the chip is removed by lightly brushing with a soft brush. The specimen is then placed in one of the cavities of a porcelain plate (tint plate with cavities for color reactions). In this way a dozen or more may be run through at the same time. The cavities are then filled with diluted bakelite (C. V. 1305).

The varnish as marketed by the manufacturers is too viscous to penetrate rocks with fine pores, so in most cases dilution is necessary. Comparative tests, run in triplicate, were made to determine which of three grades of bakelite, and which of four solvents, brought about the most rapid and complete impregnation. The following solvents were used: ether, ether and methyl alcohol (half and half), methyl alcohol, and the "thinner" manufactured by the Bakelite Corporation. In all cases varnish C. V. 1305 proved to be the best, while of the four solvents, methyl alcohol with a few drops of acetone was the most satisfactory. Methyl alcohol is to be preferred rather than ethyl alcohol because of its much lower viscosity. On diluting the bakelite with methyl alcohol, a white curdy precipitate frequently forms but quickly disappears when a little acetone is added.

The amount of solvent to be added to the varnish depends on the fineness of the rock and the length of time of impregnation. The more dilute the bakelite, the faster penetration takes place. If, however, the varnish is thinned too much, the interstices are filled mostly with solvent, which later volatilizes and leaves only a small amount of the binder in the pores. The more completely the interstitial spaces are filled with bakelite, the more rigid the particles of rock

become. It is advisable, therefore, to strike an optimum between a very dilute varnish with a short period of impregnation and a viscous varnish with a long period of soaking. In general, for loess, till, clay, etc., one volume of solvent to two volumes of bakelite has been found satisfactory, if allowed to impregnate over night under low pressure. In preparing sections of very fine laminated clays, however, it was necessary to dilute the varnish still farther.

The porcelain plate containing the rock-chips covered with diluted bakelite is placed in a desiccator to which is attached a filter pump. The pressure is reduced to about 15 to 20 cm. of mercury and the water shut off. It is advisable to place a water trap between the filter pump and the desiccator so that the water, when shut off, does not back up into the desiccator and ruin the varnish. The reduced pressure removes most of the air from the interstices of the rock and allows the diluted varnish to move into the pores by capillarity more rapidly than at ordinary pressures. Some rocks require only a few hours soaking, though in most cases it is desirable, and just as convenient, to allow them to stand over night under reduced pressure. A longer period of impregnation is sometimes necessary if the rock is very fine grained and so friable that it will not stand the preliminary grinding of a flat face before soaking. In this way a greater depth of penetration is secured, which, after baking, will permit a flat face to be ground without exceeding the depth of impregnation.

If the soaking is carried on at ordinary pressures, the time of impregnation can be reduced by starting the process with a very dilute varnish and by gradually adding more undiluted varnish, increase the concentration of the impregnator. In this way the pores are rapidly freed from air by the very dilute bakelite which later takes up the more viscous varnish. In this manner a number of sections of till were prepared with as little as three hours of impregnation. The method, however, requires considerable attention and is therefore not so desirable as an over-night period under low pressure. Impregnation under reduced pressure produces somewhat the same result, since the process begins with a dilute varnish and, as the solvent gradually evaporates, the bakelite becomes more viscous.

On removing the impregnated chips from the porcelain plate,

they are placed in an oven and the baking started. By adjusting the vents of an electric oven, the proper maximum and minimum temperatures may be produced in different parts of the oven so that only one oven is necessary for the baking process. The temperature to which the impregnated rock-chip should be subjected in baking varies with the mineral composition of the rock and with the nature of the study to be made of the thin section. If no hydrous minerals are present, the temperature may be raised to around 200° C. If hydrous minerals are present, and only the texture is to be studied, the same temperature may be used. If mineral determinations are to be made on rocks containing hydrous minerals, the temperature should not exceed 100° C. The baking should begin at about 70° C. and remain at this temperature until all of the solvent has been driven off, a period of two to six hours, depending on the amount of solvent used and the size of the rock-chip. If this first period of baking is carried on at too high a temperature, bubbling and blistering will result and the rock-chip will be disrupted by the rapid escape of the solvent. When this preliminary "set" of the varnish has been obtained, the temperature is raised to around 200° C. and kept at this temperature for two to four hours. When hydrous minerals are present it is necessary to prolong the period of baking, since the varnish does not begin to harden perceptibly until 95° to 100° C. is reached, and does not harden rapidly below 125° C. By maintaining a temperature of 100° C. for two or three days, the bakelite becomes sufficiently hardened that thin sections can be readily ground, even though complete polymerization has not been obtained.

The baking having been completed, the already flattened face is ground down on a piece of plate glass, using fine carborundum and water. Benzene is used in place of water if colloids are present. Care is necessary in grinding not to exceed the depth of impregnation so that a smooth surface is obtained. The specimen is then washed clean of all the adhering carborundum with the same medium with which it was ground.

Silts, tills, clays, etc., swell or soften when brought in contact with water. This is probably due to the forces exerted by the water in moving into the pores by capillarity, and not because of a reaction between the water and the rock-particles. It follows, then, that if the

pores are coated with some material that will hold the rock-particles rigid enough to withstand the capillary forces of the water, the rock will not be disrupted. By impregnating with bakelite and subjecting the specimen to a period of baking, the pores are coated, or even completely filled, with a substance sufficiently strong to withstand these capillary forces. Using water as a grinding medium, sections of fine varved clay, laminated clay, till, and loess have been prepared without any disintegration occurring. The impregnation must be complete, of course, or disruption will take place.

Some rocks, however, contain colloids or soluble minerals and these, when subjected to water, are disrupted by its solvent action. It is usually necessary, therefore, when making thin sections of bentonites, gumbotils, or any rock with high colloidal content, to use some medium other than water in grinding, although the varved and laminated clays (colloidal) mentioned above constitute exceptions.

Ross, after trying a number of inert liquids, concluded that either a light-grade medicinal mineral oil, or kerosene, met all the required needs. The medicinal oil can be used only when grinding by hand, since the oil clings to the revolving lap and jerks the specimen out of the hand. If kerosene is used after the rock-chip has been cemented to the glass slide, the grinding must be completed rapidly or the balsam will be dissolved and the section become loosened. In grinding down the flat face which is to be cemented to the object glass, the writer has found benzene to be more satisfactory than either petrolatum or kerosene, because it evaporates more quickly and thus allows rapid drying of the chip after it has been washed clean of abrasive. Benzene cannot be used after the specimen has been cemented to the slide because of its rapid solvent action on the balsam.

After the specimen has been thoroughly dried, it is mounted on an object glass with Canada balsam in the usual manner. The other side of the chip is then ground down as far as possible without disrupting the specimen on dry No. 1 emery cloth and is finished on a glass plate, using fine carborundum and water, or a high-grade petrolatum if colloids are present. The cover glass is mounted in the usual manner.

The procedure is not so time-consuming as would at first appear, since the longer periods consist of intervals during which no attention need be given to the rock-chips, such as soaking over night, or baking for six hours. It is apparent, therefore, that if a number of specimens be run through together and the work well organized, considerable time is saved. Most tills, silts, clays, etc., are so soft that it is generally desirable to grind by hand, even though a revolving lap is available. With a little practice one can grind sections quite thin because of the color and toughness of the bakelite. The color likewise aids in securing uniform thickness, for as one side of the section becomes thinner than another, it also becomes lighter in color. The reader may get an idea of the great toughness produced from the following: A completed section of till, approximately .03 to .04 mm. thick, was dropped about four feet on a concrete floor and, although the object glass was completely shattered, the section of till remained intact.

OUTLINE OF PROCEDURE

1. Grind down chip on dry emery cloth.
2. Brush away dust remaining on chip.
3. Place in cavity of porcelain plate.
4. Fill cavity of porcelain plate with diluted bakelite (C.V. 1305).
5. Place in desiccator over night under reduced pressure.
6. Place in oven at 70° C. for two to six hours.
7. Increase temperature to 200° C. and maintain for two to four hours. (If hydrous minerals are present, increase temperature to 100° C. and maintain for 40 to 70 hours.)
8. Grind down already flattened face, using fine carborundum and water. (If colloids are present, use benzene in place of water.)
9. Wash clean of abrasive with medium used in (8) and dry thoroughly.
10. Mount on object glass, using Canada balsam.
11. Grind down other side of chip on dry No. 1 emery cloth.
12. Finish grinding on glass plate, using fine carborundum and water. (If colloids are present, use nujol in place of water.)
13. Mount cover glass, using Canada balsam.

The foregoing detailed procedure is intended only as a rough guide in preparing thin sections of friable materials. With only slight variations here and there in the technique, thin sections have been prepared from the following materials without disturbing in the least the original structure: till, gumbotil, loess, soil, varved silt, varved clay, laminated clay, unconsolidated volcanic dust, bentonite, sandstone, shale, clay, and loose sand (with a minimum amount of pore space). Some of the varved clays were extremely fine grained and friable, the hand specimen rubbing to a fine flour and giving no gritty feel between the teeth.

TWO INTERESTING BOULDERS IN THE GLACIAL DEPOSITS OF ALBERTA

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ABSTRACT

This paper deals with the occurrence of two boulders of bituminous sandstone which have been found in the glacial deposits of central Alberta. They occur mixed with other material that was carried south by the Keewatin ice during glaciation. One boulder is a lens-shaped mass about 11 by 2½ feet. These are now more than 200 miles from the source rock and were mistaken for local oil seepages.

INTRODUCTION

During Pleistocene glaciation in Alberta quantities of rock débris were carried from the north and west and spread over the greater part of the province. Glacial deposits from the Cordilleran ice sheets are confined largely to the western part of the province in the foothills and outer ranges of the mountains, whereas the débris from the Keewatin ice sheet is spread over the greater part of the plains areas.

The glacial deposits on the plains consist largely of boulders of gneisses and schists derived from the pre-Cambrian rocks which underlie the northeastern part of Alberta and large areas to the north and east of this part of the province.

The Keewatin ice moved southward over Paleozoic and Mesozoic rocks which border the pre-Cambrian, and occasional boulders of these are included in the glacial deposits. These beds are relatively much softer than the pre-Cambrian; thus boulders of Paleozoic rocks are relatively uncommon and of Mesozoic still less common in ice-transported material.

The pre-Cambrian beds underlie such a large area that it is difficult to assign a precise locality origin to most of the boulders occurring in the glacial deposits. Similarly, it is almost equally difficult to determine the locality, except in a general way, from which the occasional boulders of Paleozoic or Cretaceous rocks may have come.

The bituminous sands of Alberta are an exception to the foregoing general statements in that they underlie a relatively small

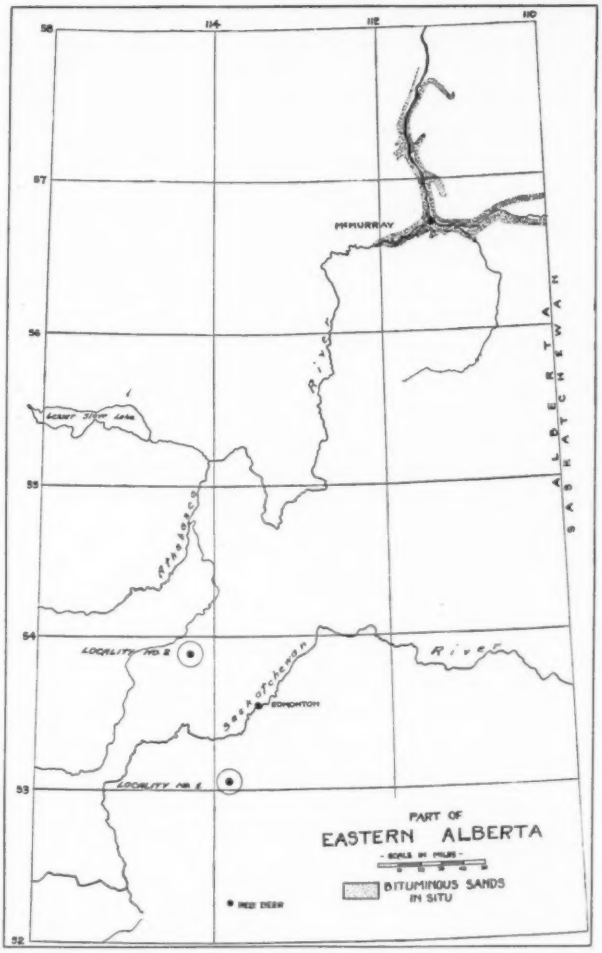

FIG. 1

area. These beds are of Cretaceous age and are sandstones which have been saturated with bitumen. They occur along Athabaska

River in the vicinity of McMurray (Fig. 1), and although they are extensive when compared to similar deposits occurring elsewhere, their distribution relative to other formations in this part of Alberta is comparatively restricted. The distribution and character of these deposits of bituminous sands are described in detail by Ells[1] and by Clark.[2] These deposits also contributed to the glacial débris, and the following is an account of two widely separated occurrences of bituminous sands in the glacial deposits.

LOCALITY 1: BONNIE GLEN

One of these occurrences is situated near Bonnie Glen rural post-office, about 16 miles west of the village of Millet, or 35 miles southwest of Edmonton (Fig. 1). The precise locality is legal subdivision 14, Sec. 14, T. 47, R. 27 W. of the fourth meridian. Here the bituminous sand is exposed at the surface and is mixed with soil and boulders of pre-Cambrian rocks. This occurrence of bituminous sand was discovered about six years ago, apparently through agricultural operations, since it was covered at first by soil and vegetation. Most of the soil has been removed from the top of the bituminous sand by prospectors and visitors gathering samples. The surface area showing pieces or blocks of bituminous sand is about 15 by 25 feet. It appears, however, that the original block has been broken apart to some extent and spread out over the surface. In places the bituminous sand is less than 5 inches thick and is underlaid by soil. Dr. K. A. Clark sampled this occurrence a few years ago and informed the writer that in all cases where the bituminous sands were dug through the soil and rock débris were found beneath.

This occurrence of bituminous sand was interpreted by some to be the result of an oil seepage, and consequently a test well was drilled near by. The derrick and equipment are still in place, but the writer has been unable to obtain definite data concerning the depth of this well, as apparently no government record has been kept. It is listed as the "Mutual Oil and Gas Company" well, and is recorded to have reached a depth of 1,100 feet.

[1] S. C. Ells, "Bituminous Sands of Northern Alberta, Mines Branch," *Pub. 632* (1926).

[2] K. A. Clark, "The Bituminous Sands of Alberta," *Sci. and Ind. Res. Coun., Alta., Rep. 18*, Part I (1927).

The writer is convinced from the nature of this occurrence of bituminous sand that it has been transported to this locality by the same agents that brought the numerous boulders of pre-Cambrian rocks to the same locality and that it does not represent an oil seepage in this district. Dr. J. A. Allan has visited this locality and is of the same opinion as the writer. Ells[1] has examined some samples of the bituminous sands from this occurrence and expresses the opinion that they are not "in place." Dr. K. A. Clark analyzed a sample taken from this occurrence by the writer and reports a bitumen content of 16.5 per cent. This value is somewhat higher than the average of the analyses of the bituminous sands at McMurray and probably indicates that only those specimens which were relatively high in bitumen could withstand long-distance ice transportation. The samples from the Bonnie Glen occurrence are strikingly similar to deposits occurring *in situ* in northern Alberta. The nearest outcrop of bedrock is over a mile from this boulder and is composed of massive sandstones of Paskapoo (Tertiary) age. There is no indication of the presence of bitumen in any of the outcrops of bed rock occurring in this district.

The geographical position of this boulder of bituminous sand is of further interest since, believing that it came from the well-known district along the Athabaska River in the vicinity of McMurray, it gives some idea of the general direction of the movement of the glacial material. Assuming that it may have come from the eastern side of these deposits, the general direction of movement would be south 25°–35° west, and if from the western side, then the general direction would be south 15°–20° west.

LOCALITY 2: NAKAMUN LAKE

A second occurrence of bituminous sand, similar to that near Bonnie Glen, is situated near Nakamun Lake about 40 miles northwest of Edmonton. The precise location is legal subdivision 14, Sec. 31, T. 56, R. 2 W. of the fifth meridian. This occurrence was first observed a few years ago when excavations were being made prior to building a house. Since then the house has been burned, but the nature of the occurrence may be readily observed in the cellar pit.

[1] S. C. Ells, *op. cit.*, p. 16 n.

The bituminous sand occurs in only the east wall of the cellar, and was cut through in the excavating. A cross-section of the boulder is lens shaped, being about 11 feet across and $2\frac{1}{2}$ feet deep at the thickest place, thinning out to a few inches at the ends. The bituminous sand is surrounded by unconsolidated recent material consisting of a mixture of residual sands and silts and fragments of pre-Cambrian rocks. There are about 2 feet of this material overlying the boulder, and about the same thickness of similar material is exposed below it. Bed rock was not exposed in the lowest part of the cellar, which is about 7 feet deep. A well has been dug within a hundred feet of the cellar, and no bituminous sands are exposed in the walls of it. This occurrence of bituminous sandstone is not "in place," but is part of the glacial deposits which are common in this and the adjacent districts. Dr. K. A. Clark analyzed a sample sent in from this locality in 1924. It was found to contain 14.7 per cent bitumen, which is close to the average of the McMurray deposits.

This was also believed by some to represent a local oil seepage, and as a result of such an interpretation drilling operations were carried on at a point about $2\frac{1}{2}$ miles to the east in the northwest quarter of Sec. 27, T. 56, R. 2 W. of the fifth meridian. The writer has been unable to secure a record of the drilling since apparently no record of it has been kept by the government.

It is almost unnecessary to state that no encouraging results were obtained either in this well or in that near Bonnie Glen. The well near Bonnie Glen started to drill approximately at the top of the Edmonton formation (Uppermost Cretaceous), whereas the well at Nakamun Lake was started at least 300 feet lower in the same formation.

If the Nakamun boulder is derived from the bituminous sand in the McMurray district, the general direction of movement in the glacial material would be south 20° west, assuming that it came from the western side, and south 40° west if from the eastern side of these deposits.

GENERAL CONSIDERATIONS

Both these occurrences have been called "boulders" since they do not belong to the underlying bed rock and apparently have been

carried a considerable distance by ice to their present respective localities. The nearest known occurrence of such beds in place are those in the McMurray district, which is at least 200 miles from the Nakamun occurrence, and 250 miles from that near Bonnie Glen. Ells[1] mentions the Nakamun occurrence with three others in the same vicinity and expresses the opinion that they are not "in place." These other occurrences are all within 25 miles of the Nakamun occurrence, and may indicate that boulders of the bituminous sand become more plentiful to the north in the direction of McMurray, where the deposits are in place. Such is to be expected, since these bituminous sands are not of sufficient hardness to withstand long-distance transportation by ice. The relatively soft character as compared to the pre-Cambrian rocks perhaps explains why they are not more prevalent in the glacial drift of Alberta. Furthermore, these two boulders of bituminous sand described above are four or five times larger than glacial boulders of other rocks found in their respective vicinities. Smaller boulders of bituminous sandstone were undoubtedly picked up by the ice, but in most instances these would be entirely destroyed before reaching such distances as Nakamun or Bonnie Glen. Consequently the size of the boulder was possibly an essential factor in determining how far south they could be transported intact.

There are perhaps many more occurrences of bituminous sand within the glacial deposits of Alberta, but since they weather comparatively readily and become covered with soil, they are not likely to be detected unless by accident, as in the case of excavating a cellar at Nakamun or cultivating land at Bonnie Glen. The two occurrences discussed above are of interest in showing how far comparatively soft material was transported by ice in Alberta. They are of interest in showing that comparatively large pieces of material were necessary to withstand the action of the transporting agent.

They are of special economic interest in that an incorrect interpretation of their mode of occurrence has in each case led to unnecessary expenditure of several thousands of dollars. Two or three days' labor with a few inexpensive tools would have proved in both cases that they were not local oil seepages.

[1] S. C. Ells, *loc. cit.*

OÖLITES OF ARTIFICIAL ORIGIN

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ABSTRACT

Oölites are described which were formed in the hot-water coil of a furnace. The oölites are composed of calcite and have nuclei of iron oxide. It is concluded that the oölites developed from the precipitation of lime carbonate around particles of iron oxide, the precipitation being due to the release of carbon dioxide as a consequence of the heating of the water. Lime carbonate was then deposited on the walls of the coil to form scale and around particles of iron oxide to form oölites.

INTRODUCTION

The problem of the origin of oölites has been more or less incidentally considered by geologists for many years and still is far from solution. Few direct observations relating to origin seem to have been made, and very little of a decisive character has attained publication. Vaughan's[1] studies of the Florida oölites, Schade's[2] observations on the growth of gallstones, and the studies made of the Carlsbad oölites are perhaps those of most importance.

As it is probable that material of oölitic structure develops under a considerable range of environments and as a consequence of several groups of combinations of factors, it does not seem to the writer to be safe to generalize with respect to their origin; and until more is known of the origin of oölitic structure, it seems to be the wisest procedure to consider each occurrence as an individual problem. Some oölites may have developed by passing from the state of an emulsion colloid as postulated by Schade and elaborated by Bucher[3] as of general application for "most if not all, oölitic and spherulitic

[1] T. W. Vaughan, *Papers from the Tortugas Laboratory, Carnegie Institution*, Vol. V (1914), pp. 49–54.

[2] H. Schade, "Zur Entstehung der Harnsteine und ähnliche konzentrich geschichteter Steine organischen und anorganischen Ursprungs," *Zeitschr., für Chemie und Industrie der Kolloide*, Vol. IV (1909), pp. 175–80; "Über Konkrementbildungen biem Vorgand der tropfigen Entmischung von Emulsionskolloiden," *Kolloidchemische Beihefte*, Vol. I (1910), pp. 375–90.

[3] W. H. Bucher, "On Öolites and Spherulites," *Jour. of Geol.*, Vol. XXVI (1918), p. 593.

grains"; others may be related to organic factors; still others may owe their origin to the gradual deposition of material about various nuclei in the manner outlined by Hochstetter,[1] and there may be several ways of origin still unthought.

The 1924 report of the Committee of Sedimentation of the United States National Research Council called attention to the formation of oölites in the hot-water coil of a furnace. It is the purpose of this article to describe these oölites, give the details of the conditions functioning during their development, and venture an explanation of the factors responsible for their origin.

CONSIDERATION OF THE FACTORS OBTAINING IN THE DEVELOPMENT OF THE OÖLITES CONSIDERED

The water circulating through the hot water coil was derived from a deep well drilled into the Dresbach formation of the Wisconsin Potsdam sandstone, the locality being on the edge of the city of Madison, Wisconsin. The well is closed at the top and the water is pumped into a pressure tank, whence it flows into the pipes supplying a house. The water has little chance of aeration. Analysis of the water of this well has not been made; but, as it is derived from a horizon very like that from which the deep wells of the closely adjacent city of Madison obtain their water, an analysis of the water of one of the wells of that city may be considered representative. The solid contents of the deep wells of Madison range from 300 to 378 parts per million. The solids from the Madison well of the Chicago, Milwaukee & St. Paul Railway, with a solid content of 356–365 parts per million for one analysis, gave as follows:

	Parts per Million
SiO_2, Al_2O_3, FeO, Fe_2O_3	Not determined
Ca	74.6
Mg	38.9
Na and K	12.6
Co_3	195.0
So_4	35.3
Cl	8.2
	364.6*

*S. Weidman and A. R. Schultz, "Water Supplies of Wisconsin," *Bull. 35* (1915), p. 300.

[1] H. V. Hockstetter, *Karlsbad, seine geognostische Verhaltnisse und seine Quellen* (1856).

The coil in which the oölites formed is in a furnace and is connected with a water tank at the bottom and the top. The water enters the pipe under pressures which range from 55 pounds to atmospheric pressure. During times of use of hot water the circulation is fairly rapid; at other times circulation is slow, as it is due entirely to the heat differential. The coil is merely a pipe which crosses and returns in horizontal position along one side of the furnace in close U-shaped arrangement. It was installed in 1920 and was first opened in 1924 to ascertain if it required cleaning, as there seemed to be some retardation of circulation. Also at this time a water softener was installed so that subsequently soft water circulated through the coil.

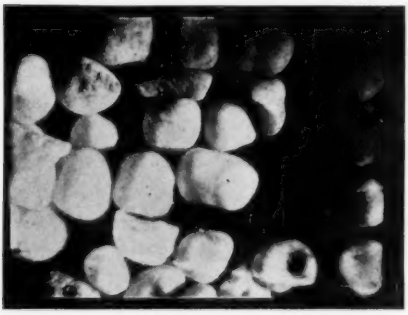

FIG. 1.—Microphotograph of the oölites, ten times natural size.

When the coil was taken out for cleaning, it was found to contain at least a thousand small spherical particles. Some of these were broken and were found to have concentric oölitic structure. The walls of the pipe had little scale, and that present had a botryoidal appearance, with some of the spherical particles adhering. The oölites range in diameter from approximately one-fourth to three-fourths millimeter, and are of various degrees of sphericity. So far as sectioned they seem to have small particles of iron oxide as nuclei and the ultimate shapes of the oölites seem to have been determined by the shapes of the nuclei. Compared to the volume of an oölite, that of the nucleus in those sectioned constitutes a very small part, perhaps a thousandth of less part for an oölite three-fourths millimeter in diameter. The variations in the shapes of the oölites are shown in Figure 1. The structures of the oölites seem invariably to be concentric, and the number of laminae ranges from five or less to more than twenty, the fewer number of laminae occurring in the smaller oölites. Similar banding occurs in the scale on the walls.

Nothing resembling radial structure has been observed in an oölite as a whole, or in any part of the laminae of an oölite. The absence of radial structure in the laminae, however, is by no means certain, as each lamina is very thin, and it may be that radial structure is present in some instances. It has not been seen, however, under high magnification in a single instance. The laminae are outlined by slight differences in color, and ordinarily are not visible to the eye. The internal structure of the oölites is shown in Figure 2. The dimension of an oölite is almost directly proportional to the number of laminae, that is, an oölite with five laminae has its diameter about one-fourth the length of the diameter of an oölite with twenty laminae.

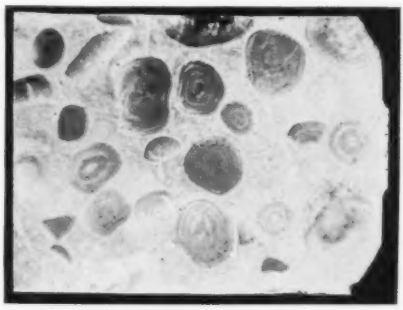

FIG. 2.—Microphotograph of a thin section of the oölites, ten times natural size.

The composition of the oölites seems to be almost entirely carbonate of lime. It is possible that some magnesium and iron carbonate may be present. The composing mineral is calcite and not aragonite.

The coil was again opened on July 26, 1927, but no oölites were present. Evidently they do not form in the softened waters.

ORIGIN OF THE OÖLITES

As the oölites were not observed during their development, it is not possible to state positively the manner of their formation. However, certain factors are known, and these afford the basis for certain rather definite conclusions. The known factors are as follows: The water entering the coil is not of high temperature; its solid content consists in large part of carbonates which probably are in true solution; there is some iron in the water, but the quantity is small and it may be expected to be in the form of the carbonate or sulphate; most of the carbonate is calcium; there is considerable range in circulation and there is little chance for aeration. The quantity of carbonate in the water is so small that the quantity precipitated

at any one time is too small to permit its accumulation into particles as large as those of any of the oölites. The scale is like the oölites in composition and structure, with the slight difference that the structure of the scale is parallel to the wall of the tube, instead of concentric as in the oölites. The same factors are responsible for the oölites as for the scale. There is little probability that the carbonate when precipitated accumulated into small bodies of gel, but if such was the case the same must be held for the scale of the walls. The latter is improbable. The carbonate was aggregated about small particles of iron oxide, or material which at present seems to be iron oxide.

It seems extremely probable that the inciting cause of the precipitation was the warming of the water, leading to the breaking up of the bicarbonate of calcium and probably also magnesium and iron. As a consequence the calcium carbonate was precipitated from solution and deposited upon the walls of the coil, where it formed scale, and also upon any small particles in suspension or lying upon the floor of the coil. As the particles were subject to some movement, deposition upon all sides resulted. The particles in suspension or on the floor are thought to have been iron filings left in the pipe, pieces of rust from the pipe, or particles of iron oxide precipitated from the water itself.

As the laminae are about one-fortieth to one-fiftieth millimeters thick, it seems obvious that the rate of deposition of the composing material must have been extremely slow, and as a consequence there was little possibility of the oölites becoming cemented to the wall and thus becoming a part of the scale. That this sometimes occurred is shown by a few oölites found adhering to the walls. There seems little probability that each oölite originated as an aggregate and then became concentric in its structure through diffusion of matter from its center, but if such a view is entertained for the oölites it should also be maintained for the scale. This seems absurd.

It would be vain to apply the method of origin outlined to all oölites, as the conditions and the environment are unlike those present in nature. On the other hand, it is considered extremely probable that oölites may not uncommonly develop in nature in accordance with the method outlined. This is a qualitative statement; nothing is assumed from the quantitative point of view.

A DIABASE CHERT BRECCIA[1]

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ABSTRACT

A diabase dike containing chert fragments linearly distributed in irregular manner is regarded as a graphic expression of progressive faulting and pulsational intrusion.

Students of Ontario pre-Cambrian geology learn early to regard diabase dikes as potential evidence of large or small faults. This does not mean that every diabase dike is the locus of a fault; but it is a common experience to trace formations to such a dike and, on crossing the dike, to find the formations offset, sometimes a great deal. Such a dike was encountered during the past field season, and in a rather typical way it marked the locus of a fault.

The occurrence was found in the township of Cunningham, a few miles to the northeast of Ridout Station on the Canadian Pacific Railway. In this township to the east of Isaiah Creek and 2 miles south of its mouth is a rather extensive iron formation which strikes almost due east near the creek and dips quite steeply. All the formations are strongly faulted, and some of the faults show considerable displacement.

Some faults, notably those which occur within the iron formation, are marked by breccias. The breccias which occur in the iron formation consist of angular fragments of cherty iron formation, usually cemented by vein quartz, but sometimes by finely ground iron formation itself, which is thoroughly indurated. Usually the chert fragments are small, measuring 6 inches or less in diameter; but larger fragments are not uncommon. The brecciated masses are irregular in shape, but those which occur on the margins of the iron-formation masses usually have a strong linear trend. This is particularly true of a chert breccia which was found in repeated outcrop along the south margin of the iron formation referred to east of Isaiah Creek. This breccia marks the eastern recognized part of a

[1] Published with the permission of the director of the Geological Survey of Canada.

prominent fault. Near the creek to the west the fault is marked by a diabase dike.

For nearly half a mile along the south margin of this iron formation, commencing slightly more than a quarter mile east of the creek, the chert breccia and the diabase dike referred to, join. To the north of the diabase dike is iron formation, and to the south is greenstone schist and granite. Interest centers on the northernmost 50 to 60 feet of the dike, where it contains chert fragments in an interesting manner. To the south of this, the diabase is essentially free from

FIG. 1.—Occurrence of chert fragments in diabase

inclusions over its extent of several hundred feet. This diabase chert breccia is quite unique in the writer's experience. It is believed to suggest a graphic expression of a typical manner of igneous intrusion and faulting and to be the exceptional case in which the evidence is preserved.

The chert fragments occur in lenticular areas of various sizes from a few feet to 100 or 200 feet long, and from a few inches in thickness to 9 or 10 feet. The largest single lens was found on the south side of the first hill east of the creek. Between the various lenses the diabase contains only a few chert fragments, and to the south the lenses of inclusions disappear. The accompanying photograph is of an outcrop which was well exposed, and illustrates the

typical occurrence. It can be seen in the photograph that the fragments which are frequently elongate parallel to their bedding planes have no uniform orientation. The lenses, the dike, and the main chert breccia all strike in general parallel to the iron-formation mass. The chert fragments making up the lenses are a mixture of various aspects of the iron formation; they are not of one type of chert, such as a shattered single block.

Many explanations of the phenomena observed in these outcrops have occurred to the writer and have been suggested to him. But there is only one to which serious objections have not been found.

The statement is frequently found in the literature that faulting in general, or in a specific area, is known to take place not in one movement but in a series of movements separated by periods of stability. According to this well-founded conception, the progressively cumulating stresses are satisfied periodically by release. In other words, movement along a particular fault plane is an interrupted or periodic one, the final effect being the resultant of all movements. Usually there is little or no evidence of the steps through which the end was attained.

Again the common genetic connection between periods of faulting and periods of igneous intrusion has been made repeatedly. For this the evidence is commonly much stronger and more manifest. I believe that the diabase and chert breccia described here are direct evidence of these features.

The manner of local faulting and intrusion as it appears to the writer with respect to these features is as follows: Along a prominent east-west fracture plane, connected, I believe genetically, with local intrusion of a diabase mass, there was movement of the north block relative to the south block. The brittle iron formation cut by this fracture plane produced on movement an amount of iron-formation fragments distributed along the vertical fracture. This was followed by the intrusion of a thin sheet of diabase magma, which appears to have varied in thickness and to have solidified after engulfing, but not dissolving, the chert fragments. This formed a series of lenses of diabase well charged with fragments and distributed along the fracture plane. But since there are many lenses side by side, it seems necessary to allow more than one such thin intrusion. The writer

believes there was a series of intrusions which probably alternated with further movement and brecciation. All these intrusions were irregular, and only some had access to the supply of fractured iron formation. The large width of chert-free diabase to the south of the breccia lenses may have been one larger in-welling of magma, or several smaller ones which did not have access to the supply of fragments.

It seems impossible to me to avoid the main conclusion that the igneous material was introduced in a pulsational way. If it is assumed that large fragments of iron formation were included in the intruded dike and later shattered in some way to produce the present breccia lenses, the mechanics of the procedure seems unreasonable in view of the regular distribution of the breccia masses and the irregular orientation of the fragments within the lenses. Also, the lenses themselves, as pointed out, do not consist of a single type of cherty iron formation, but of various aspects of it.

Again it has been suggested that if a single intrusion had access to a supply of fragments, these fragments would collect in eddies, thereby forming islands of fragments in the diabase. The marked lenticular shape of the islands and the scattered fragments linearly distributed beyond the islands militate against this view.

The interpretation offered here is the writer's. It is not intended to imply that his understanding of these outcrops is complete. Other interpretations are undoubtedly possible. But the outcrops seem to contain critical information which is worthy of attention.

REVIEWS

Einführung in die allgemeine Kohlenpetrographie. By R. POTONIÉ. Berlin: Gebrüder Borntraeger, 1924. Pp. x+85; text figures, 80.

This very interesting book deals with the origin and the classification of the coals, their structures, and the substances of the plant and animal bodies which contribute to the building up of the coal seams. R. Potonié's book supplements the well-known book by his father, Henry Potonié, on the origin of coals and of the other mineral fuels. As would be expected, R. Potonié follows closely in the footsteps of his father's theories and adds many additional facts supporting them. Also the chemical quality of the coal receives a good deal of attention, and it is, whenever possible, explained out of the constituent vegetation of the original coal bed.

Of special interest in R. Potonié's book are the many practical suggestions for laboratory technique with regard to examination of the coal. The book is well illustrated and should prove extremely helpful to any student of this interesting subject, be he a geologist, a botanist, or a chemist.

A. C. N.

Die Braunkohlen Deutschlands. By KURT PIETZSCH. Berlin: Gebrüder Borntraeger, 1925. Pp. xii+488, pls. 20, figs. 105.

The book forms the first volume of the third series of the *Handbuch der Geologie und Bodenschätze Deutschlands* (*Textbook of the Geology and Mineral Deposits of Germany*), edited by Erich Krenkel.

When Germany lost, permanently or temporarily, part of her bituminous coal deposits in Upper Silesia and in the Sarre region, attention was turned toward the enormous lignite deposits as a solution of the domestic fuel problem. The production of lignite was greatly increased, and as a consequence the geologic and technical study of lignite became very important.

This book is divided into three parts, of which the first deals with the general geologic conditions of the German lignites. This includes a discussion of the physical and chemical properties of lignites, the general structure of lignite seams, their origin, and the floral and faunal condi-

tions, as well as the climate of the periods in which lignites were deposited. They are all of Tertiary age.

The second part describes the different lignite deposits. They are arranged in a South German, a West German, a Central German, and an East German group. Also the total amount of the lignite deposits of Germany is estimated, and the author's figure is slightly above 22,000,000,000 metric tons.

The third part of the book gives a survey of mining methods applied to the German lignite deposits and the preparation of the coal for the trade, especially briquetting. Here the chemical treatment of coal, especially hydrogenation and coking, are discussed. A bibliography on the subject appears at the end of the book.

Of general geologic interest is a chapter on the general genetic conditions of the German lignite deposits. Here we find discussed whether the lignite seams have originated *in situ* or whether the plant material which formed these seams had been carried from distant places, and what are the criteria for recognizing each type of coal-seam formation. The author comes to the conclusion that coal seams of both types exist. The geologic and climatic conditions for the formation of lignite seams are also discussed, as are the structural changes which took place in the seam subsequent to its formation.

Comparatively short is the treatment of the flora and fauna of the lignite seams, although occasional references to these topics are scattered throughout the book.

The chemical changes which the plant material underwent in changing to coal are also considered.

Considered as a whole, Pietzsch's book on the lignites of Germany will prove of interest to the economic geologist as well as to the structural geologist, and also, to a considerable degree, to the paleontologist.

A. C. N.

The Properties of Silica. By ROBERT B. SOSMAN. Pp. 856. New York: Chemical Catalog Co., Inc., 1927. $12.50.

This work is an exceptionally complete handbook on all phases of the subject, excluding the hydrated forms. The treatment is very lucid, a simple exposition of the general subject preceding the various detailed descriptions; in fact, as stated by the author, "the book has tended to take the form of a textbook on the properties of substances in the crystalline and glassy states, with silica as the leading example." The style is

personal—as easy as an informal conversation—and yet brief. The broad philosophical point of view is one that could well be emulated in many other scientific books.

The forty-seven chapters in the work cover the ground suggested in the title in comprehensive fashion. There are numerous references to the literature, and a good number of line figures and tables. No short review can adequately note the full scope of the treatise. For the physical geologist or the mineralogist-crystallographer the book will not only serve to bring his perhaps neglected physical-chemistry and allied subjects up to date, but a section on geological applications (stressing quartz thermometry) will be of particular interest to the former, while the latter will find a part on symmetry and structure and a section on silica minerals and rocks, as well as much other instructive material. The work is apparently remarkably free from errors, though novaculite is said to be produced by contact metamorphism, which origin hardly fits the Arkansas occurrences.

The *Journal* congratulates Dr. Sosman, the American Chemical Society, and the Geophysical Laboratory of the Carnegie Institution of Washington on this exhaustive publication.

D. J. F.

The Geology of the Virginia Triassic. By JOSEPH KENT ROBERTS. Charlottesville, Virginia, 1928. Pp. xii+205, figs. 19, pls. 32.

This is a thesis presented for the Doctor's degree at Johns Hopkins University and published both as such and as *Bulletin 29* of the Virginia Geological Survey. The area involved is divided into some seven separate basins that, with a few small outliers, aggregate about 1,350 square miles. Four summers were devoted to the field work, and later the igneous rocks, originally planned for Dr. Thomas L. Watson to work, were, after his death, studied by Dr. Roberts.

The several areas are on the piedmont plateau and occupy narrow down-faulted troughs that have since suffered from deep erosion and peneplanation. The author, however, inclines to the view that these several areas were probably never connected, but developed as separate basins. The two more easterly areas, the Richmond and Farmville, stand in contrast to the remaining, more westerly areas, in that they developed under mesophytic, or swampy, conditions while the others were characterized by dryer continental conditions. It is not believed by the author, however, that the red beds indicate that the climate was that of a desert, but rather that it was as moist and warm, perhaps, as southern Georgia today.

The deep weathering of the generally level piedmont surface has made detailed work difficult, especially in tracing fault structures. The rocks are mainly sedimentary, but, like the rest of the eastern Triassic, include important areas of igneous rock. The sediments include some half-dozen types of border, basal conglomerates, the Manassas red sandstone, and the Bull Run shales. The igneous rocks are diabases. They cover an area of more than 250 square miles and include stocks, dikes, and a few intrusive sheets. There are a number of very excellent photomicrographs of the several kinds of diabase.

The general structure of the several basins is a westward-dipping monocline bounded by a western marginal fault. The persistent westward dips that vary from a few degrees up to 45° give an appearance of enormous thickness to the Triassic, which the author believes is merely apparent and due to repeated faulting. It is thought very doubtful if the original thickness of the sediments reached 2,000 feet.

The coal of the Richmond basin is the most important economic resource of the Virginia Triassic. Its structure and quality are discussed at some length. The sandstone, trap, and Potomac marble are also described, the trap being the only one of present importance. A section is given to a summary of our knowledge of the animal and plant life, and one to a sketch of environmental conditions of Triassic time.

The report is a careful, comprehensive, and well-balanced piece of work, and the views expressed on controversial points—such as desert climate and thickness of beds—appeal to the reviewer as sane and reasonable. Sedimentary petrography receives interesting treatment. The photographs are unusually good and there is a carefully annotated bibliography.

L. C. GLENN

Geologic Investigations in Western Kansas. By N. W. BASS. State Geological Survey of Kansas, *Bull. 11*. Lawrence, 1926. Pp. 96; maps.

The main part of this report consists of summaries of the geology of Ellis and Hamilton counties, Kansas. The larger part of these counties is covered by Cretaceous rocks, in which Bass has observed many structural features due to differential settling over old unconformities. These features are usually noticed only when connected with faulting, which is very common in the area. A review of the geologic structure of the Dakota sandstone in western Kansas and a short summary of the structure and areal distribution of the salt beds are given in separate chapters.

R. C. Moore appends a brief notice on a large sink in Sharon County.

C. C. B.